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SECTION IV

GEOLOGICAL SCIENCES

INCLUDING MINERALOGY

THIRD SERIES—VOLUME LIV—SECTION IV

JUNE, 1960

OTTAWA

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TRANSACTIONS OF THE ROYAL SOCIETY OF CANADA
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SECTION FOUR

PRESIDENTIAL ADDRESS

On Turbidites

F. FITZ OSBORNE, F.R.S.C.

ABSTRACT

Turbidity currents of high density are known to have developed in oceanic basins, but their properties are not fully understood. The theory that similar currents have transported some coarser clastic sediments now seen in marine sequences is moderately to well supported. The study of a fluxoturbidite and a quartzite turbidite suggests that most of the outstanding lithologic characteristics of the turbidites are determined by the characteristics of the original sediments that caused the turbidity current.

THE possibility that some rocks were deposited from turbidity currents has been seriously considered by the sedimentary petrologist during the last two decades. Many sedimentary sequences have been inferred to contain turbidites; and perhaps the novelty of the concept has led to its application to rocks for which it is not valid. A principal difficulty in the application of the concept is that, although the evolution of deposits of coarser clastic sediments is easily observed along shores and streams, the locus of formation of and deposition from turbidity currents of high density is outside the region of direct observation, and accordingly a bed can be judged a turbidite only by inference.

A turbidity current forms where a mass of fluid contains suspended sediment. The density of the suspension is greater than that of the fluid alone, and the suspension is thus capable of flowing under a body of the fluid. The fluid is commonly water, and the formation of underflows of fresh water with a small tenor of silt or clay has been observed and analysed. Turbidity currents of high density in marine water are more interesting to geologists than those of clay in fresh water. Much of the pioneer work on currents of high density was done by Kuenen, but recently Heezen (9) has presented a summary of the problem of such currents.

Grand Banks Turbidity Current of 1929

The underflow associated with the Grand Banks earthquake of 1929 has received a prominent place because the interruption of transatlantic cable services gave almost laboratory conditions to determine the time of passage of the underflow. Geologists and geophysicists have used diverse models in

calculating the properties of the underflow, and therefore results are different, particularly for the proportion of silt in and the thickness of the underflow. Plapp and Mitchell (11, p. 989) calculated that in the distal part, approximately 450 nautical miles from the top of the continental slope, the underflow was 470 metres thick and contained a fraction of silt comparable to that in very turbid rivers. However, Kuenen calculated the thickness as 120 metres, and Stonely gave it as 3.6 metres. The density of the suspensions would accordingly be greater than that calculated for a thicker flow. What is perhaps the most significant parameter to the geologist, and one about which there can be no disagreement, is the gradient of the ocean floor. For the first part of the flow, that is, to 250 nautical miles from the top of the continental slope, the average gradient is about 1.5 per cent. Here the velocity reached more than fifty knots. From 250 to 450 nautical miles—the segment in which the velocity declined—the gradient is 0.05 per cent. These low gradients are important in discussing the formation of turbidites in lithified sequences.

Properties of Suspensions

The behaviour of suspended solids in liquids is of interest in several branches of engineering. The movement of sediment in streams, flumes, or pipes under the influence of gravity has been much discussed, but complete mathematical solutions have not yet been evolved (4). The pumping of suspensions through pipes presents analogous problems concerning the properties of mixtures. Some properties of the suspensions in motion show deviations, some of which are velocity dependant, from the properties of a simple fluid to an extent that suggests that these or similar phenomena may allow a turbidity current to form and to retain its identity as an underflow.

Mineral dressing is concerned with the movement of particles or grains through fluid, but it is more preoccupied with the grains themselves, and hence data from mineral dressing are useful in considering the origin of sedimentary rocks. Furthermore, because of the necessity of using a minimum quantity of water and the smallest possible apparatus, many of the suspensions are of high density and thus provide information on the properties of the inferred turbidity currents. Many of the data in the ensuing summary are from Gaudin (6). The velocity of movement of small rigid spheres through a large body of still fluid is given by Stokes' equation; that for larger spheres is given by a modification of Newton's equation. Non-spherical bodies of the same material fall more slowly than spheres, and the larger the mass of the particle the greater the disparity. A sphere of quartz 1.2 cm. in diameter falls at the same rate in water as a disk 100 cm. in diameter and 3.5 cm. thick. If the volume of the solid is a substantial part of the volume of the liquid, conditions of hindered settling obtain and the velocity of setting is reduced. Experiments show that the velocity is about 10 per cent of free settling if γ, the volume fraction of solid in suspension,

is 30 per cent, and with γ's of 10 and 20 per cent the velocities are reduced to 43 and 23 per cent of the free fall velocity respectively.

The equation for settling commonly gives the velocity approached by the falling grain. Small grains reach this velocity in almost infinitesimal time and distance. However, for larger spheres time and distance are appreciable. A quartz sphere 2 cm. diameter in water reaches 95 per cent of its terminal velocity after travelling 70 cm.

The foregoing summary does not explain the formation of turbidity currents, but reveals enough of the properties of suspensions to show that some mechanism may exist to endow them with their peculiar behaviour.

Turbidites

One of the geologist's tasks is to recognize turbidites in a sequence of rocks. It is obvious that such recognition is hampered by the absence of full understanding of how they are formed. Ten Haaf (8) has laid down criteria for turbidites, but he recognizes their inadequacy for he says (p. 217): "as to the properties of turbidites there is no actualistic evidence." Like most workers, ten Haaf emphasizes the massiveness of the beds of turbidites and the presence of one kind of graded bedding. The fragments of many turbidites are poorly sorted, and the grains which settle faster tend to be at the bottom of the bed. Thus, with grains of the same substance of the same sphericity, the larger grains are at the bottom. In a sequence of beds, however, some may show grading but a nearby bed of the same composition may not. The grading is in effect a classification that develops as turbulence decreases, and is comparable to the arrangement that would develop in a thickener as feed and agitating devices were cut off. The operation of thickeners and classifiers shows that the classification of fine particles is feasible in only dilute suspensions, where γ is 1 to 2 per cent, but suspensions of particles the size of sand allow classification, with γ up to 12 to 18 per cent (6, p. 205). These figures suggest an upper limit of density for the turbidity currents which give rise to turbidites with marked graded bedding. This upper limit raises the problem of the driving force for the current. The driving force is the difference between the density of the suspension and of the clear liquid. For quartz and water it is 1.67 γ.

In 1936 Daly suggested that the velocity of an underflow be described by a modification of Chézy's equation, into which effective density, D, that is, the difference of density between suspension and nearby liquid, is introduced. The formula is $v = C\ (R\ S\ D)$. R stands for the hydraulic radius, which for a wide underflow is half its thickness; S is the slope of the bottom, which is presumed to be parallel to the top of the underflow, and is expressed as a sine; C represents a constant determined in part by R and S and also by the character of the material in suspension and by velocity. This equation has been used in several models proposed for turbidity underflows, but its application is of doubtful validity except for turbidity currents of low density

in which a steady state is achieved and, even for these, some experiments (7) suggest that the formula might not fully describe the behaviour of the underflow.

For turbidity flows of high density the applicability of the equation is even more uncertain. A steady state is unlikely to be achieved, and the probability that water is added to the suspension prevents Bernoulli's continuity condition being met.

Manning's formula can be derived from the Chézy formula and modified by the introduction of the effective density, but the effective density is a function of γ and, for quartz and water, $D = 1.67\,\gamma$. Also $R = T/2$. The modified equation is then,

$$v = \frac{K}{n} T^{2/3} S^{\frac{1}{2}} \gamma^{\frac{1}{2}}.$$

The slope is S as in the Chézy equation, K is a new constant, n is Manning's friction factor. The equation in this form shows the pronounced effect on velocity of the thickness of the underflow. To attain a large velocity it is obvious that γ should be large. Kuenen considered that the effective density is about 0.6, which would be a γ of about 35 per cent for quartz and water. However, if the mineral dressing data are valid, such a γ seems to be too great to allow effective grading of the deposit, unless, as velocity falls, dilution increases.

It is obvious that turbulence in the underflow is necessary to maintain the suspension. The turbulence does not extend into the overlying boundary layers and so does not cause extra fluid to become involved in the underflow apparently because some mechanism not now understood prevents this. Various methods have been used to compute the work done in maintaining turbulence, but data from mineral dressing indicate that the work required may not be great. The power required to maintain the suspension of an aggregate of minerals in a tank thickener is equivalent to a drop of the mass of sediment of less than one-eighth of an inch per minute. It is the practice, however, to use a motor supplying many times this energy on a thickener.

Clays in Turbidites

Clays have received special attention in discussions of turbidites because of their importance in currents of low density and their presence in greywackes, many of which are inferred to be turbidites. Govier (7) has given the results of measurement of suspensions of "Delaware clay" in a concentric cylinder viscosimeter and has demonstrated that, even at concentrations as low as 5 per cent by weight, the suspension behaves as a pseudo-plastic with some characteristics of a Bingham plastic. H. A. Einstein (5) used an Ostwald water viscosimeter to measure the properties of suspensions at different densities and degrees of flocculation. He states that the degree of flocculation determines the plasticity of the aggregate. In a personal com-

munication to the author, Govier states that during his experiments no evidence of flocculation was observed. This is obviously an intriguing field for investigation.

Clays have another effect, which is not well understood. It was long ago observed in mineral dressing that the addition of a small amount of clay to a suspension of sand in a launder results in an increased velocity of flow. Comparable observations have been made on flow in natural channels and in pipes. Two explanations have been put forward: the clay decreases the friction (Manning's n) on the wet periphery, or the clay tends to decrease turbulence, thereby decreasing the resistance to flow. Perhaps both processes operate.

A geologist has little difficulty in visualizing the formation of a current of low density where a clay- or silt-laden stream enters standing water. However, the genesis of high density currents with clay and particularly the way in which clay is acquired with sand by turbidity currents are more difficult to explain. A body of unlithified sediment or largely unlithified sediment in standing water becomes unstable, slips downslope, and acquires an increased tenor of water of the order of 50 per cent of the aggregate by volume. The sediments thus form a subaqueous landslide which passes with further increase of water into a turbidity current. The deposits formed in the transition are known as fluxoturbidites and have characteristics of both landslides and turbidites.

Most turbidites have considerable clay in the matrix, and some have flakes or pebbles of shale. It is obvious that some clay is made available by the attrition which produced the pebbles, but, in some turbidites, the shale masses are sub-rounded and substantial contributions of clay from attrition seem improbable. The cohesiveness of clay and its passage into shales largely by compaction suggest that massive beds of clay are unlikely to become dispersed as clay as a result of deformation accompanying sliding to form the matrix clay. It is interesting to note that the marine Leda clays along the St. Lawrence River almost invariably give a Swedish-type landslide, and the slide blocks of clay in the zone of strong wave action, along the estuary of St. Lawrence for example, persist for many months.

Turbidity currents also incorporate clays deposited on the sea floor across which they pass. Newly deposited and flocculated clay may have as little as 20 per cent solid by volume. The solid ratio increases by exudation as the clays compact and develop cohesiveness. The freshly deposited clay may well be picked up by the currents, but small recumbent folds and sandstone—or rather greywacke—dykes in shales beneath turbidites show that the clays were substantially consolidated when the flow occurred. The possibility that some clay is swept up by the current cannot be denied, but some turbidites in shale sequences have no matrix clay.

Another possible explanation of the presence of clay is that the original sediment which was contributed to the underflow had matrix clay. This is probably the source for much clay.

The Charny Fluxoturbidite

In 1953 I gave a description of some of the sandstones[1] of the Charny, Lower Cambrian, formation near Logan's line at Quebec Bridge. These sandstones and small-pebble conglomerates are interbedded with siltstones and red shales, are lenticular in habit, and show other characteristics, including graded bedding, of fluxoturbidites. Concretions up to a foot across cemented by carbonate occur in a small-pebble conglomerate and preserve some of the original sediments forming the underflow. These concretions are slightly bedded and are composed of poorly sorted subangular grains comparable in size and roundness to many of the grains that make up all or some of the present sandstone beds. It is inferred that the carbonate has replaced the clay matrix of the original rock. The beds that contain the concretions also have slabs of silty dolomite one-and-one-half inches thick and two feet wide. A few pebbles of this dolomite can be recognized in some beds, but they are far from abundant. A friable greenish sandstone with argillaceous cement forms disk-shaped pebbles in some beds.

Most descriptions of the Charny sandstones mention the occurrence of well-rounded pebbles of quartz. Similarly rounded pebbles of black shale have not been mentioned as often, but they are present in many beds. One lenticular bed is composed of 60 per cent quartz pebbles from 1.2 cm. to .2 cm. in diameter with 40 per cent shale pebbles of similar size. The roundness of the pebbles and granules is noteworthy and extends to even the smaller size which are subangular in the concretions. These rounded pebbles are probably products of inshore erosion and sorting and are substantially younger than the sandstone represented by the concretions. They slide with the older sediments and various proportions of them become incorporated in the underflows.

The fluxoturbidites near Quebec Bridge show that not only clays but older fluxoturbidite beds also were eroded. The relations between the beds are in marked contrast with those seen in normal turbidites, such as those in the Charny formation farther from Logan's line. If the evidence at Quebec Bridge is valid the underflow can, during a transitional stage between a subaqueous landslide and a turbidity current, pick up material from the bottom over which it flows, although at the same time it picks up considerable water.

The Kamouraska Turbidites

I have inferred that the Kamouraska quartzites of the Quebec group are turbidites, and, because the data from them complement those from the Charny formation, I am presenting a brief discussion of them. The Kamouraska quartzites, which are not older than late Cambrian and the probably

[1]I have retained Logan's name for the coarser clastics near the Quebec Bridge. It is possible to find a wide variety of clastic rocks for which such names as greywacke, grits, and small-pebble conglomerate can be used.

early Ordovician, outcrop for about 100 miles, with a breadth up to twelve miles on the south shore of the St. Lawrence River beginning thirty miles downriver from Quebec. The area in which they outcrop has not been mapped recently, but there are apparently at least three zones. The thickest zone is about 500 feet and all are intercalated with grey or black very thin-bedded shales and silstones.

The quartzites are in massive beds more than three feet thick and are devoid of crossbedding, ripple-marks, scour channels, or other evidence of deposition in shallow water. In places very thin and persistent beds of shale occur between sandstones, but in most places a thick sandstone bed floors a similar thick bed. The quartzites are composed of almost spherical quartz grains from 1 to 2 mm. in diameter cemented by quartz. A few grains of feldspar and calcareous oolites of size and sphericity comparable to those of the quartz grains occur. Clay does not form any of the matrix, but flattened angular chips of shale are found in some of the sandstones. The chips are several times the diameter of the sand grains, and those from the top of a bed are large enough to settle equally with the quartz grains, allowing for the effect of shape. In some beds the larger chips of shale tend to be concentrated near the base, and even more obvious graded bedding is apparent in some beds with quartz pebbles and cobbles at the base. In the lower part of some thick beds, boulders of limestone two feet or more in diameter are abundant.

Field evidence seems to indicate that the Kamouraska quartzites are turbidites. I suggest that older sedimentary rocks were subject to working and sorting by waves, and beds of quartz sand and clays were deposited along a coast that became intermittently unstable and allowed the accumulated material to slide downslope. The moving material with added water formed turbidity currents that ultimately deposited the massive beds which are now quartzite. In spite of the fact that some quartzites rest on shale, the absence of clay in the matrices of the quartzites is noteworthy.

Conclusions

In foregoing sections I have discussed the Charny fluxoturbidites and also the occurrence of pure quartzites deposited by turbidity currents. In a very great measure the two discussions complement each other. They indicate that, except in the transitional zone between subaqueous slide and turbidity current, little material is picked up by the current from the floor of the sea and that the composition of the turbidite is determined by the composition of the original sediments that were contributed to the underflow. This conclusion is of some interest in trying to detect in the field the source from which the turbidity current derived its sediment. For example, it is possible that the crossbedded greywacke of the "Timiskaming series" illustrated by Wilson (14, plate 4) is a relic of a formation that contributed sediment to turbidity currents.

The turbidity currents obviously moved downslope away from the source. The sediment of the Charny fluxoturbidites moved generally southeast, that is, away from the Shield, and the direction of movement is the same in some lithologically similar turbidites ninety miles downriver from Quebec. In both cases movement was across the basin of deposition. However, the axial line of such a basin may plunge, and movement along the axis may result. Descriptions of the turbidites show that such longitudinal motion is common, and in many instances the underflow moves in different directions to form succeeding beds. The nature of the Kamouraska quartzite makes it difficult to determine its direction of motion, although distribution of the quartzite suggests but does not prove that longitudinal movement was present if not dominant.

In conclusion, turbidity currents offer a plausible explanation for at least some beds of sedimentary rocks, but proof of such a medium of transportation demands careful study not only of the field relationships but also of the petrography of the deposits. Undoubtedly some deposits have been erroneously attributed to turbidity currents. In discussions of turbidity currents of high density, it is particularly interesting to note the place taken by what should be a rock name, "greywacke." Authors have used turbidity currents to explain the emplacement of any rock given this name, but more commonly the name has been used for rocks believed to be turbidites (2), despite the fact that many of the rocks do not conform to the rock definition.

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*The bibliography of turbidites is so long that references to some recent works only are given together with references to a bibliography of them (1). The paper by Heezen (9) has many references on turbidites. The references to stream transport and to movement of suspension through pipes are those papers that I have found most useful.

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SECTION FOUR

Observations on the Biogeochemistry of Lead in Canada

HARRY V. WARREN, F.R.S.C., AND ROBERT E. DELAVAULT

ABSTRACT

Workers in other countries have found lead in most vegetation which they have analysed, usually in the range of from 10 to 100 parts per million (p.p.m.) in ash.

In Western Canadian vegetation we have noted significant variations in the lead content of various species of trees and lesser plants growing in close association: obviously vegetal matter differs widely in its capacity to absorb lead.

Probably the most far-reaching result of our investigations has been to learn that some gasoline exhaust fumes are responsible for higher concentrations of lead in vegetation close to highways than any other cause yet encountered: concentrations of over 1000 p.p.m. in ash have been noted in vegetation from several localities. Only vegetation growing close to significant lead mineralization shows comparable concentrations.

GEOCHEMISTRY is constantly providing new methods of assisting in the search for buried ore bodies. Soil, stream sediment, and water testing have all proved their worth in mineral exploration. We believe that there are localities where biogeochemistry may be used effectively and, with this idea in mind, we have made a start in determining normal and anomalous lead contents in some of the trees and lesser plants most frequently encountered in southern Canada.

Although the lead content of soils has been widely studied in some countries, we can find no record of any systematic investigation being made in Canada nor do any biogeochemical studies on lead appear to have been published. In Finland, Lounamaa (13) has not only reported on the lead content of many rocks and soils, but has also made a semi-quantitative spectrographic survey of the trace elements, including lead, in lichens, mosses, ferns, conifers, deciduous trees and shrubs, dwarf shrubs, grasses, and herbs. We have found Lounamaa's work so valuable that we make no apology for frequently referring to it.

A majority of the trees and shrubs which we have analysed carry less than 1 p.p.m. of lead in dry plant material, and samples containing more than 4 p.p.m. usually may be considered anomalous. The fact that in many areas, there are specimens which run 10 p.p.m. or more, corresponding to an ash content of upwards of 500 p.p.m. suggests that the lead content of vegetation may be of interest to agriculturists and medical men, as well as to geologists.

Lead in Rocks

Before investigating the lead content of vegetation, it seems pertinent to consider briefly the lead content of the earth's crust, of several soils, and of some of the types of rock most likely to contain an anomalously high lead content.

Both Goldschmidt (8, p. 398) and Vinogradov (20, p. 155) accept 16 p.p.m. as being the average content of lead in the upper lithosphere. However, as might be expected, lead occurs in widely differing amounts in the various types of rock occurring in the earth's crust. Geochemical principles suggest at least three types of rock that are apt to contain amounts of lead greater than normal.

Lead, as a lithophile element, occurs in a large number of rock-forming minerals. Borchert (1) and Wedepohl (23), have reported on the lead content of a number of minerals and rocks. Thanks to its ionic radius (Pb^{2+} 1.32 Å), lead replaces potassium, and, in certain minerals, even calcium (1.06 Å). The presence of lead in the potash feldspars of magmatic rocks has recently been well demonstrated in Norway by Heier and Taylor (9), and in the United States by Slawson and Mackowski (18). The work of the Geological Survey of Canada (10) strongly suggests that at least some of the granitic rocks of Nova Scotia carry abnormally high amounts of lead.

We can report on 112 calcareous rocks selected from various areas in southern Ontario. The lead content of the great majority of these rocks runs from 1 to 5 p.p.m. However, "shaly dolomites" contain noticeably larger amounts, usually from 15 to 20 p.p.m., and one dolomitic formation carries from 35 to 200 p.p.m. of lead.

Goldschmidt (8, p. 402), also draws attention to the fact that in hot arid climates the oxide of quadrivalent lead is found: this has a rutile structure (Pb^{4+} 0.89Å). Goldschmidt notes that important deposits of lead have been formed where porous material occurred in a region of arid climate under oxidizing conditions: the new red sandstones of Mechernich in the Eifel and early Cambrian or late Precambrian detrital sandstones in northwest Sweden are cited as examples.

Lastly, we must remember that lead is a chalcophile element, and that it is from the mineral galena that virtually all our lead is obtained.

Lead in Soils

Surprisingly little is known about the distribution of lead in soils in spite of the fact that Swaine (19, p. 86) records that in 1915 its presence was reported in some soils of the western United States. Until 1946, nobody appears to have tried systematically to relate the lead content of soils to the lead content of the parent rocks. Then in Scotland, Mitchell (14) and Swaine (19) compared the lead content of various soils derived from different rocks. Other investigations followed, but, until Lounamaa (13, pp.

34–59), nobody seems to have tackled the problem of relating the lead content of soils to the lead content of the parent rocks. Lounamaa, after reviewing the soil and rock relations for all the other elements which he studied, wrote of lead (13, p. 59): "Lead differs from all the other aforedescribed elements in that the soil samples from the different outcrops invariably contain more of it than do the rocks. The strong enrichment of lead into the soils overlying silicic rocks is quite remarkable."

On the basis of admittedly scanty evidence, Vinogradov (20, p. 155) points to the A. horizon as being one of lead accumulation. Much, but not all, of the information we have been able to collect in Canada supports this suggestion. Goldschmidt (8, p. 402) writes: "Many investigations have been made on the occurrence of lead in various soils, but little is known about the fixation of lead, and about its migration between various horizons."

If we exclude soils derived from rocks which contain abnormally high quantities of lead, the data available at present suggest that normal soils may be expected to contain from 0.5 to 5.0 p.p.m. of lead. We have found soils in Canada with abnormally high lead contents which may be caused by lead mineralization where the lead in the related soil may run more than 1 per cent, or by abnormal parent material which may give rise to soils carrying from 100 to 200 p.p.m. or more of lead. Extraneous contamination must also be considered and, in this connection, contamination from two sources merit mention, exhaust fumes from some gasolines, and the products of some orchard sprays. The former, presumably, are charged with tetraethyl lead, and the latter with lead arsenate. In British Columbia, where orchards some ten years ago were sprayed with lead, we have found soils containing from 40 to 100 p.p.m. of lead, in contrast to soils which were not subjected to these sprays, and which carried less than 1 p.p.m. Likewise, some soils collected from within one hundred yards of an arterial highway in British Columbia ran less than 1 p.p.m. when analysed by our usual prospecting procedures. It may be presumed that this low figure was caused by the volatility and solubility of the lead from the exhaust fumes. More appropriate analytical methods actually showed from less than 1 to around 100 p.p.m. lead in these soils.

On the basis of only sixty analyses of Canadian soils, we can do no more than suggest that all soils contain some lead, usually between .5 and 5 p.p.m. Soils over some rocks which contain high quantities of lead may run from 25 to 300 p.p.m. Soils may be contaminated by mine waste, by orchard sprays, such as lead arsenate, and by the exhaust fumes of tetraethyl gasoline. Where significant, and sometimes insignificant, mineralization occurs, soils carrying 1 per cent, and even more, lead may be encountered.

Lead in Vegetation

In order to determine the normal lead content of vegetation, samples were collected from nineteen different localities in southern British Columbia,

TABLE

TYPICAL LEAD CONTENT IN SOME REPRESENTATIVE SPECIES OF TREES AND LESSER PLANTS IN

Suite No.	Locality and district (author responsible for background geology)	Nature of Mantle and underlying rock	Family Genus *Pinus* Rainfall	(*a*) *monticola* (*b*) *ponderosa* (*c*) *contorta*
1	St. Mary's River at junction with Redding Creek E. Kootenay, Rice (15)	Fluvioglacial sands and gravel. Aldridge metamorphosed sediments	40″	
2	Hell Roaring Creek near junction with St. Mary's River, E. Kootenay, Rice (15)	Fluvioglacial sands and gravel. Aldridge metamorphosed sediments	30″	
3	St. Mary's River and Kootenay River junction, E. Kootenay, Leech (11)	Glacial silts and till plus St. Mary's River silts. Bedrock unknown	15″	
4	Kootenay River near Wasa E. Kootenay, Leech (11)	Glacial silts and till. Bedrock unknown	15″	
5	Mather Creek near Highway No. 95, E. Kootenay, Leech (11)	Glacial drift, Creston metamorphosed sediments	20″	*b* 2 (110)
6	Duhamel Creek near Kootenay Lake E. Kootenay, Little (12)	Poorly developed soil over fluvioglacial debris	25″	*b* 3 (98)
7	Richter Pass, S. Okanagan, Cairnes (2)	Thin mantle augite syenite	10″	*b* .5 (29)
8	Richter Pass, S. Okanagan, Cairnes (2)	Thin mantle granodiorite	10″	*b* .5 (27)
9	East of Alberni near Horne Lake Road W. Vancouver Island, Clapp (5)	Heavy mantle of glacial drift. Sandstone and shale	80″	
10	Malahat, above Mill Bay E. Vancouver Island, Clapp (6)	Light mantle of glacial drift. Granodiorite	40″	*a* 4 (210)
11	Metchosin Peninsula S.W. Vancouver Island, Cook (7)	Light mantle of glacial drift. Augite gabbro	40″	*c* 1 (54)

where geological mapping by the Geological Survey of Canada has provided sufficient data to enable geochemical evaluation of each site. Areas were selected where the rainfall varied from ten inches to eighty inches annually, where rocks were covered by a mantle a few inches to several hundred feet thick, and were made up of material which ranged from gravel to glacial clays, stream silts, and even some poorly developed soils.

Obviously, the results which follow must be considered to be merely exploratory to more detailed efforts. Not only have there been no publications in Canada dealing with the lead content of vegetation, but it was neces-

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OF SOUTHERN BRITISH COLUMBIA. LEAD IN P.P.M. OF DRY PLANT, AND OF ASH (SHOWN BRACKETS)

Pinaceae		*Cupressaceae*	*Salicaceae*	*Betulaceae*		*Rosaceae*	*Aceraceae*
Pseudotsuga	*Tsuga*	*Thuya*	*Salix*	*Alnus*	*Betula*	*Prunus*	*Acer*
Menziesii	*hetero-phylla*	*plicata*	(*a*) *scouler-iana* (*b*) *argo-phylla*	(*a*) *tenui-folia* (*b*) *rubra* (*c*) *sinuata*	(*a*) *fonti-nalis* (*b*) *papy-rifera*	(*a*) *demissa* (*b*) *emargi-nata*	(*a*) *macro-phyllum* (*b*) *glabrum*
	1 (78)		*b* .5 (16)	*c* .3 (16)			*b* .5 (20)
1 (43)			*b* .3 (12)	*a* .5 (26)	*b* 1 (50)		
			b 3 (110)	*a* .9 (43)	*a* 2 (83)	*a* 3 (48)	
			b .2 (7)		*a* .8 (50)	*a* 3 (80)	
4 (150)							
8 (29)	4 (140)	2 (55)	*a* 2 (63)		*b* 1 (48)	*b* 2 (100)	*b* 1 (45)
1 (35)						*b* .7 (8)	*b* .3 (8)
.4 (14)					*a* .4 (27)		
.4 (29)		.8 (23)		*b* .4 (15)		*b* .7 (33)	*b* .4 (13)
18 (900)	45 (2600)	4 (110)	*a* 5 (140)	*b* 2 (93)			*a* .5 (17)
1 (60)	3 (180)	1 (74)	*a* .3 (10)	*b* .7 (35)			*a* .3 (8)

sary also to try out and settle on some reasonably simple and relatively foolproof method of analysis: X-ray fluorescence, and arc spectroscopy proved unacceptable, and were abandoned in favour of a chemical attack which is outlined below.

Samples were collected during September 1959 and, in order to save space, only the results obtained from better known species reasonably well represented at eleven different sites are tabulated in Table I. The results obtained from other sites merely confirm the findings from the reported localities, although, as has been found to be the case with other metals,

some species appear to eschew, and others effectively to concentrate, lead. This fact suggests that, if lead is harmful when consumed by humans in organo-metallic form, studies should be made of the varying amounts of lead taken up by food plants.

Discussion of Results

When the suites were collected, only No. 3 was expected to exhibit any evidence of an anomaly. Water and silt testing at the Kootenay River, just below its junction with the St. Mary River, showed abnormally high zinc values. Actually, the trees from this site were reported to contain abnormal amounts of zinc: *Salix argophylla* (willow) contained 1.6 per cent zinc in the ash. However, as will be noted by looking at Table I, the lead contents from this suite did not appear abnormal, particularly when compared with those obtained from corresponding genera from suites Nos. 5, 6, and 10. It was then realized that both No. 5 and No. 10 suites had unwittingly been collected from vegetation which had been exposed to fumes from automobile exhausts, the samples having been taken from within a couple of hundred feet of a highway. Suite No. 6 from Duhamel Creek, near Nelson, showed erratic results, which although we are now prepared to believe could have been caused by the vagaries of automobile fumes, because the samples were taken within a quarter of a mile of a main highway, may represent the typical scatter pattern which we have found occasionally when rock from a mineralogical area is dispersed by a glacier. This latter possibility must be investigated further.

When all the samples taken at sites 5 and 10 were considered, the results left little doubt as to the efficacy with which the lead from burnt gasoline could enter vegetation. However, thirty more samples were taken of various conifers growing within one hundred yards of a highway in the

TABLE II

Normal Background

Species	Organ	No. of samples	p.p.m. Pb. in ash	
			Range	Average
Picea mariana	2nd yr. Stems	6	8–37	21
Pseudotsuga menziesii	2nd yr. Stems	4	14–60	38
Abies grandis	2nd yr. Stems	2	23–56	39
Pinus pondorosa	2nd yr. Stems	2	27–29	28
Thuya plicata	2nd yr. Stems	3	23–74	51
Betula glandulosa	2nd yr. Stems	7	21–53	32
Betula fontinalis	2nd yr. Stems	3	13–50	30
Alnus rubra	2nd yr. Stems	2	15–35	25
Salix sp.	1st yr. Stems	3	11–15	13
Amelandier alnifolia	1st yr. Stems	4	10–22	13
Shepherdia canadensis	1st yr. Stems	2	33–46	39
Ledum sp.	2nd yr. Stems	32	7–42	17
Symphoricarpos albus	1st yr. Stems	3	8–25	19
Prunus demissa	2nd yr. Stems	1	8	8
Prunus emarginata	2nd yr. Stems	1	33	33

TABLE III

MODERATELY ANOMALOUS BACKGROUND

Species	Organ	No. of samples	p.p.m. Pb. in ash	
			Range	Average
Picea mariana	2nd yr. Stems	20	20–130	66
Larix occidentalis	2nd yr. Stems	2	120–140	130
Pseudotsuga menziesii	1st yr. Stems	3	200–260	220
Pinus monticola	2nd yr. Stems	2	110–210	165
Tsuga heterophylla	2nd yr. Stems	2	140–180	160
Thuya plicata	2nd yr. Stems	1	110	110
Betula fontinalis	2nd yr. Stems	1	83	83
Beula glandulosa	2nd yr. Stems	40	56–1200	250
Salix scouleriana	2nd yr. Stems	3	63–140	104
Alnus rubra	2nd yr. Stems	1	93	93
Shepherdia canadensis	2nd yr. Stems	1	250	250
Prunus demissa	2nd yr. Stems	1	48	48
Prunus emarginata	2nd yr. Stems	1	100	100

Vancouver area. The results were conclusive, and appear in Table IV. One tree actually had an ash content of 0.4 per cent lead, and many ran more than 0.1 per cent lead.

In Tables II, III, and IV, samples whose lead comes from exhaust fumes are mixed with samples whose lead comes directly or indirectly from lead mineralization: in considering only the lead content, no distinction can be made as to the origin of the lead. Fortunately, most lead mineralization is accompanied by a little zinc, and, in practice, it is frequently possible to distinguish an anomaly caused by mineralization from one caused by exhaust fumes: the former will have a zinc content that is higher than normal.

The results in these tables are presented in three groups corresponding to vegetation which, in the light of our present knowledge, may be considered normal, moderately anomalous, and strongly anomalous. These tables indicate clearly that, in due course, biogeochemistry should undoubtedly prove to be a useful tool in any search for lead mineralization. Our results compare closely with those obtained by Lounamaa: although we were concerned with different species, the genera were the same, and the

TABLE IV

STRONGLY ANOMALOUS BACKGROUND

Species	Organ	No. of samples	p.p.m. Pb. in ash	
			Range	Average
Pseudotsuga menziesii	2nd yr. Stems	10	510–2200	1200
" "	2nd yr. Needles	10	290–1200	600
Thuya plicata	2nd yr. Stems	7	630–3100	1600
" "	2nd yr. Needles	7	190–940	500
Tsuga heterophylla	2nd yr. Stems	10	270–1100	650
" "	2nd yr. Needles	10	120–1100	600
Abies grandis	2nd yr. Stems	4	350–1100	800
" "	2nd yr. Needles	4	130–420	300
Betula glandulosa	2nd yr. Stems	27	270–2500	1300

parallelism of our two sets of results is highly gratifying. Incidentally, we are in agreement with Lounamaa on the lead content of leaves and stems: we both found that stems carry much more lead than leaves.

As we have mentioned earlier, lead tends to accumulate in a particular soil horizon, usually the A_1. Forest litter and peat may also accumulate lead. Because of variations in the thickness and development of soil horizons, and because of difficulties inherent in identifying the components and correctly sampling forest litter and peat, the sampling of specific organs of any particular species of tree may well provide the most satisfactory method for seeking out areas anomalous in lead.

The lead content of vegetation may be of use not only to prospectors but also to agriculturists and foresters. A colleague H. L. Cannon (3) who has been working on soils and vegetal matter in two widely separated limestone areas, one in Maryland, and one in New York, reports that, although the underlying rocks contain 30 p.p.m. of lead, cultivated gardens over these rocks contain an average of 318 p.p.m. lead, and reach a maximum of 1000 p.p.m. Furthermore, the ash of vegetables and fruit growing on these gardens runs up to 500 or 700 p.p.m. lead.

There is a wealth of information concerning the iron, manganese, potash, calcium, magnesium, phosphorus, copper, zinc, cobalt, and molybdenum content of rocks and soils. A study of the silver, lead, and mercury content of soils and vegetation might well prove useful to the mineral industry and to agriculturists.

Analytical Method I

Plants are ashed at a low-red heat, and the resulting ash is dissolved in warm hydrochloric acid (21). Most soils have been roasted at low-red heat and refluxed in boiling 1 N sulphuric acid (22). To avoid losing lead which has been introduced by artificial fumes, some soils have been attacked by twice their weight of distilled nitric acid, which is evaporated, then by half their weight of lead-free aqua regia, which is also evaporated. The residue is taken up by a few drops of 6 N hydrochloric acid, water is added, and the solution filtered on an acid-washed filter.

Lead determinations have been carried out by a dithizone mixed colour method (17).

A known volume of acid solution, usually representing 0.5 grams of plant, is passed into a 50 or 25 ml. glass-stoppered cylinder. If this solution is very acid, it should be brought to near-neutrality by lead-free ammonia. Any ferric iron which may be present is reduced by adding 1 mg. of ascorbic acid for each ml. of solution, or a few milligrams in excess of what is needed. The solution is made to pH 8.5 to 9, by adding 2 ml. (or half its volume) of a 50 per cent ammonium citrate solution containing enough ammonia to make its pH about 9. Copper and zinc are complexed by adding 10 per cent in volume of 10 per cent potassium cyanide.

A 15-mgm.-per-litre solution of dithizone in chloroform is added in successive increments of 0.5, 0.5, 1.0, 1.0, 1.0 ml. until a purple colour persists on shaking for ten seconds.

Analytical Method II

The unknown which is to be determined is then compared with a series of standards made up as follows:

millilitres of dithizone	0.5	1	1	1	2	2	3	3	4
gammas[1] of lead	0.2	0.5	0.8	1	1.5	2	3	5	8

For high samples one may take a smaller aliquot or add enough dithizone to match the colour obtained with 4 ml. matched with 8 gammas of lead: in this latter case each millilitre used represents two gammas of lead contained in the solution being tested.

Conclusions

Published data and personal inquiries indicate that in Canada there has been virtually no research on the biogeochemistry of lead. However, vegetal matter shows the presence of anomalous quantities of lead in soils and other mantle material as well as the more widely studied elements, such as copper, zinc, molybdenum, and cobalt.

What may be referred to as "industrial salting" provides a more widespread and serious problem than was anticipated. Fortunately this "industrial salting," although it may provide a serious problem for society generally, can usually be recognized, and need provide no real obstacle for those who wish to use biogeochemistry for universal exploration.

Acknowledgments

This work represents a truly gratifying co-operative effort, and it is not practical to list all those individuals who have assisted in collecting various samples across Canada. The Departments of Mines and of Agriculture, both of the Dominion and of several provinces, have helped far beyond the needs of courtesy or formality. The actual funds needed for special equipment, collecting, and analyses, have come, in part, from the Defence Research Board of Canada, the Geological Survey of Canada, and the Dominion Department of Health and Welfare, with the co-operation of the Provincial Department of Health Services and Hospital Insurance. Kennco Explorations (Canada) Limited, Consolidated Mining and Smelting Company (Canada) Limited, and United Keno Hills Mines Limited have all aided in this project, the last-named by providing many soil and

[1]A gamma (γ) = one thousandth of one milligram.

plant samples from over or near their ore bodies in the Yukon Territory. We thank sincerely all the above organizations for their valuable assistance. We are also indebted to Mr. J. Davidson, who identified all the botanical samples and assisted in their collection.

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TRANSACTIONS OF THE ROYAL SOCIETY OF CANADA
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SECTION FOUR

A New *Freboldiceras* from the Canadian Arctic

P. S. WARREN, F.R.S.C., AND C. R. STELCK

ABSTRACT

A new species of the Lower Cretaceous Albian ammonite *Freboldiceras* is described from Hume River, Northwest Territories, from 174 feet below *Subarctoplites belli*. *Freboldiceras* apparently evolved directly from *Cleoniceras*.

IN a recent paper, Dr. R. W. Imlay of the United States Geological Survey described a new ammonite genus *Freboldiceras* from the Albian of Alaska. Collections examined by the writers in 1959 from the Canadian sub-Arctic disclosed a new species belonging to this genus. As with the genus type, it is associated with *Beudanticeras glabrum* (Whiteaves), which ranges through the lower part of the Middle Albian stage. In a recent paper (10), the writers recorded *Freboldiceras* only from Alaska and the Peace River area of British Columbia. This new discovery gives the genus boreal rather than Pacific distribution, and also shows its stratigraphic position to be below *Subarctoplites belli* (McLearn). On Hume River near Potter Creek, Latitude 65°23′ N, Longitude 129°57′ W, *Freboldiceras imlayi* n.sp. occurs 225 feet above the base of the Cretaceous section in the Sans Sault formation. *Subarctoplites* cf. *belli* (McLearn) occurs 399 feet above the base of the section and at 635 feet above the base there is a closely ribbed undescribed species of *Subarctoplites* (ct. 1171, U. of A.) which appears to be transitional between *S. mcconnelli* (Whiteaves) and a primitive *Gastroplites* with a rounded venter. The assumption made by Imlay (6, p. 181) that *Freboldiceras* occurs below *Lemuroceras belli* McLearn is now substantiated.

ORDER: *Ammonoidea*
GENUS: *Freboldiceras* Imlay, 1959

Freboldiceras imlayi Warren & Stelck, new species. Plate I, Figs. 6–12.

Description. Shell compressed, discoidal, whorls sub-ovate in section, higher than wide, thickest towards the umbilicus, embracing about two-thirds. Flanks gently convex in umbilicolateral portion but tapering to a rounded venter, which is more narrowly rounded in the earlier whorls. Umbilicus about one-quarter of diameter, walls steeply inclined, rounding evenly into the flanks. Living chamber at least three-quarters of last whorl.

The ornamentation of the septate portion of the shell consists mostly of prominent, flexuous, regularly spaced primary ribs overlain by fine striae.

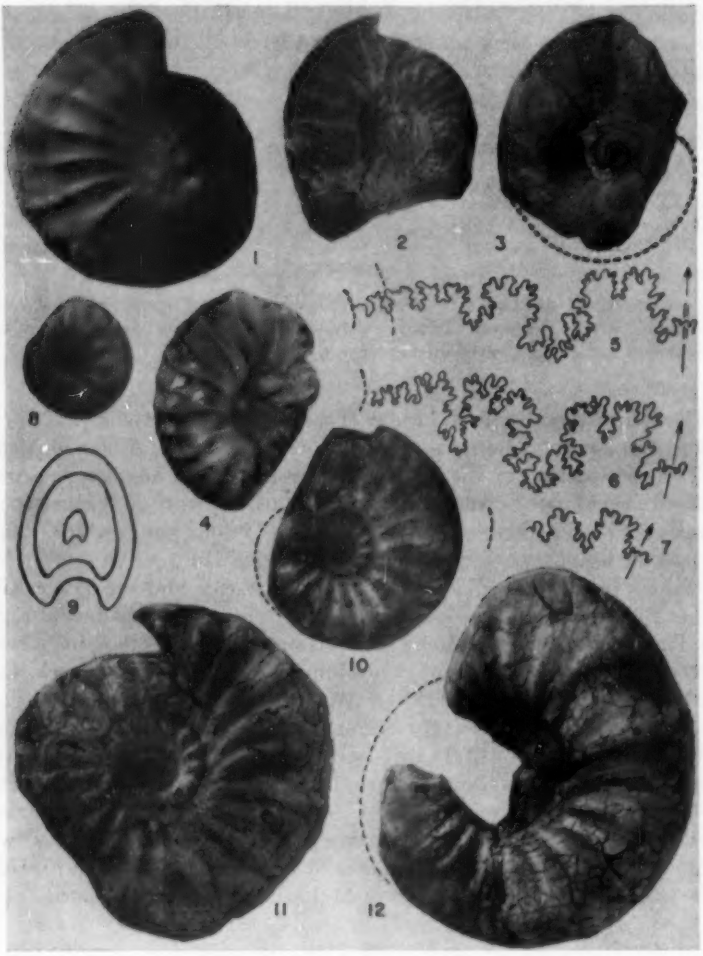

PLATE I: *Freboldiceras, Subarctoplites* AND *Beudanticeras*

Figs. 1 and 2. *Subarctoplites irenense* (McLearn) (×1). 1: Ct. 1162, U. of A. from upper part of Moosebar shale, Hudson Hope, British Columbia. 2: Ct. 1163, U. of A. from middle Buckinghorse shale, Chischa River, British Columbia. Fig. 3. *Beudanticeras* cf. *glabrum* (Whiteaves) (×1). Ct. 1164, U. of A. from Hume River, N.W.T. in association with *Freboldiceras imlayi* N.S. Figs. 4 and 5. *Subarctoplites belli* (McLearn). 4: Ct. 1166, U. of A. (×1) from Hume River, N.W.T., 174 feet stratigraphically above *Freboldiceras imlayi* N.S. 5: Ct. 1160a, U. of A. (×4) from upper part of Loon River formation, Peace River, Alberta. External suture, whorl height 15 mm. Figs. 6–12. *Freboldiceras imlayi* Warren & Stelck, N.S. 6, 9–12, from Hume River, N.W.T., 225 feet above the base of the Cretaceous section;

Occasional constrictions may appear on some specimens. Primary ribs begin high on umbilical wall on the earlier whorls but reach the umbilical seam on the living chamber. They incline forward and become swollen near the umbilical edge in the penultimate whorl but inflation is delayed to a latero-umbilical position on the living chamber. Ribs curve forward at the ventral shoulder and in the earlier whorls occasional weakened expression of ribs cross the venter. Fine striae, parallel to the ribs, ornament the surface of the shell, about five to each intercostal space. The periphery of the internal mould seems to be undulating, but where the shell is preserved the undulations are reduced. The adult body chamber shows reduction of ribbing on the periphery and on the ventrolateral portions and a thinning of the rib on the umbilicolateral portion. There are 21 primary ribs on the ultimate whorl, 17 on the penultimate, and 14 on the antipenultimate whorl.

The suture line is simple, although retaining a definite frilled appearance. It has a broad first lateral saddle slightly wider than the first lateral lobe and about equal to the second lateral saddle. First lateral lobe is unequally bifurcate and slightly deeper than the ventral lobe. Second lateral lobe is very narrow.

Dimensions of holotype and paratypes:

	Ct. 1168	Ct. 1169	Ct. 1167
Maximum diameter	7.8 cm.	8.6 cm.	5.0 cm.
Minimum diameter	6.5 cm.	7.8 cm.	4.7 cm.
Height of whorl, beginning of living chamber	3.0 cm.	2.4 cm.	2.0 cm.
Width of whorl, beginning of living chamber	2.0 cm.	2.0 cm.	1.4 cm.
Umbilical diameter	2.3 cm.	2.6 cm.	1.5 cm.

Paratype Ct. 1169 shows rapid expansion from a whorl height 2.4 cm. at the beginning of the living chamber to 3.6 cm. at one-quarter whorl later.

Type. Holotype, University of Alberta Geological Museum, Ct. 1168
Paratypes, University of Alberta Geological Museum, Ct. 1169, 1167
Figured spec., University of Alberta Geological Museum, Ct. 1170.

Occurrence. Sans Sault formation: Ct. 1167, 1168, 1169, from 225 feet above base, Hume River, Lat. 65°23′N, Long. 129°57′W, N.W.T.
Ct. 1170, float, Lower Anderson River.

Remarks. Freboldiceras imlayi differs from the genus type *F. singulare* Imlay in its greater size, in its retention of stronger lateral ribbing to the

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6: suture (×2.5), Ct. 1165, whorl height 2.4 cm.; 9: whorl sections at heights 3.5, 2.2, and 0.7 cm.; 10: Ct. 1167 (×1) paratype; 11: Ct. 1168 (×1) holotype; 12: Ct. 1169 (×1) paratype. 7–8: from Anderson River, N.W.T., Ct. 1170, U. of A.; 7: (×7), external suture at whorl height 0.5 cm.; 8: (×2).

end of the living chamber, and in its more advanced dissection of the suture. The new species lacks definite constrictions present in the genus type. Ct. 1170 apparently represents the inner whorls of *F. imlayi* and the suture at whorl height of 0.5 cm. is practically identical with that shown for *F. singulare* (Imlay, Pl. 30, Fig. 5). The suture of specimen Ct. 1169 at whorl height 2.5 cm. is somewhat reminiscent of *Cymahoplites kerenskianus* Bogoslowsky (2, Pl. XI, Fig. 1c). It differs from the sutures shown for *Subarctoplites* by Warren and Stelck (10, Pl. II, Figs. 1–4) in having a much broader second lateral saddle.

The ribs on the living chamber in the new species are more oblique to the umbilical wall than in "*Lemuroceras*" *irenense* McLearn (7, Pl. V, Fig. 5). The latter also lacks the strong umbilical bullae present on the phragmacone of *Freboldiceras*.

From the comparison of the earlier whorls the writers are led to the conclusion that *Freboldiceras imlayi* and *F. singulare* were derived directly from *Cleoniceratid* stock without an intervening *Subarctoplites* stage; whereas "*Lemuroceras*" *irenense* McLearn is actually a *Subarctoplites* that develops *Freboldiceras*-like ornament on the ultimate whorl and is therefore derived indirectly from *Cleoniceras*. This explains the anomalous position of "*Lemuroceras*" *irenense* occurring with, and somewhat above, *Subarctoplites belli* whereas the similarly ornamented *Freboldiceras* occurs stratigraphically below *S. belli*.

Another apparent homeomorph of *Freboldiceras imlayi* is found in *Kossmatella whitneyi* (Gabb) as illustrated by Anderson (1, Pl. 21, Figs. 1, 2), but the more rounded whorl section and the arborescent suture of the latter serve to distinguish the two forms.

GENUS: *Beudanticeras* Hitzel, 1905

Beudanticeras cf. *affine* (Whiteaves). Plate I, Fig. 3.

Beudanticeras cf. *affine* (Whiteaves) McLearn (1945), Geol. Surv. Canada, Paper 44–17, Pl. V, Fig. 6.

Remarks. Two specimens of this species were collected with the holotype of *Freboldiceras imlayi*. The figured disk (Ct. 1164, U. of A.) is a fragmentary inner whorl at the diameter of 55 mm. The unfigured disk (Ct. 1165, U. of A.) is a fragmentary outer whorl including part of the living chamber at diameter of 165 mm. This association would suggest that occurrences of *Freboldiceras* in Alaska and Canada are contemporaneous, or nearly so.

GENUS: *Subarctoplites* Casey, 1954

Subarctoplites belli (McLearn). Plate I, Figs. 4, 5.

Lemuroceras belli McLearn 1945, Geol. Surv. Canada, Paper 44-17, Pl. III, Figs. 17, 18. Also Appendix, p. 2.

Subarctoplites belli (McLearn) Warren and Stelck 1959, Trans. Roy. Soc. Canada, vol. 53, Section IV, Pl. II, Figs. 4, 9–12.

Remarks. The illustrated half disk of *S. belli* (Ct. 1166, U. of A.) is from the same locality as the holotype *Freboldiceras imlayi* but from a horizon 174 feet stratigraphically higher. We can find no significant difference between this specimen and topotypes of *S. belli*. The suture of *S. belli* shown in Figure 5 is taken from a topotype of the species (Ct. 1160a, U. of A.). It is inserted to illustrate the similarity of the suture of *S. belli* to that of *Freboldiceras imlayi*.

Subarctoplites irenense (McLearn). Plate I, Figs. 1, 2.

Lemuroceras irenense McLearn 1945, Geol. Surv. Canada, Paper 44-17, Pl. V, Fig. 5. Also Appendix, p. 2.

Hypotypes. Plate I, Fig. 1, Ct. 1162, U. of A.: living chamber from the Peace River, in the upper part of the Moosebar shale, Hudson Hope, B.C. Living chamber only.
Plate I, Fig. 2, Ct. 1163, U. of A.: from Chischa Creek, B.C., in the Buckinghorse shale, 450 feet above the base, occurring with *Subarctoplites* cf. *belli*.

Remarks. Subarctoplites irenense has been tentatively compared with *Freboldiceras* by Imlay (6) and by Warren and Stelck (10). The inner whorls of *S. irenense*, however, pass through a stage in which both primary and secondary ribbing are equally developed on the flanks (see Fig. 2) with reduction of the primaries on the umbilical margin. *Freboldiceras imlayi* retains strong latero-umbilical expression of the primaries throughout. For this reason the writers would retain *Lemuroceras irenense* McLearn within the genus *Subarctoplites* rather than assign it to the genus *Freboldiceras*. It does provide an excellent example of convergent homeomorphy within this group of ammonites.

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TRANSACTIONS OF THE ROYAL SOCIETY OF CANADA
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SECTION FOUR

Deformation of Soils by Glacier Ice and the Influence of Pore Pressures and Permafrost

W. H. MATHEWS, F.R.S.C., AND J. R. MACKAY, F.R.S.C.

ABSTRACT

Deformation of unconsolidated material by glacier ice has been observed in the Western Canadian Arctic and has been reported in other glaciated regions. Deformation seems commonly related to an actively moving ice margin where the surface slope of the ice was probably the steepest. High pore pressures, which probably developed in the soil, may have facilitated thrusting by reduction in the shear strength of the soils (cf. 10). The presence of permafrost, its aggradation and degradation on land and under the sea, is considered in relation to shearing strength and the development of pore pressures.

DISTURBED unconsolidated strata, some as young as Pleistocene, are known in glaciated areas throughout the northern hemisphere from Leningrad in Russia to Herschel Island in northwestern Canada. The disturbed beds have generally been explained as push moraines thrust in front of ice-sheets, as beds crumpled, folded, and faulted by the overriding action of ice against a topographic obstruction, and as slices of ground stripped off and transported by an advancing glacier. The processes, so far as we are aware, have not been seen in action, with the exception of the formation of small frontal push moraines. Indeed, the opportunities for observation are almost completely precluded for processes which are restricted to a subglacial environment. However, the recent and notable advances in the fields of glaciology and of permafrost studies, coupled with the already substantial science of soil mechanics, make it possible to offer suggestions on the process of subglacial thrusting which can be applied to examples of deformed sediments. It is the purpose of this paper to emphasize that development of high fluid pressures in unconsolidated sediments beneath a glacier may facilitate ice thrusting by reducing the shear strength of the soils.

Distribution and Nature of Disturbed Strata

Although disturbed strata are of wide distribution in glaciated areas, there is considerable evidence to support the view that deformation commonly occurred near the margin of an actively moving lobe. In the case of push moraines and advancing ice pressing against a topographic obstruction,

deformation would, of necessity, have been associated with ice front conditions. For the other types of deformation, more or less sharp curvature of the folds and fault traces, the association of disturbed beds with end moraines, their trend normal to glacial fluting, and their common occurrence near the limits of glaciation, all indicate deformation near glacier terminuses. Specific or implied mention of the relation between deformation and the ice margin have been made by numerous authors, such as Byers (1) for Claybank, Saskatchewan; Flint (3) for eastern Dakota; Fuller (4, p. 207) for Long Island; Horberg (8, p. 322) for the Lethbridge, Alberta, area; Jessen (11, p. 140) for Denmark; Mackay (13) for the western Arctic coast of Canada; and Rutten (22) for the Netherlands and Western Germany.

Most observers of ice-thrust features have stressed their superficial nature. Ice-thrust features in unconsolidated material rarely exceed one hundred metres in thickness. Beds may be tilted, folded, thrust faulted, and overturned. Slabs of unconsolidated material, more than fifty metres in thickness, have frequently been broken into large fragments and stacked in uniformly inclined positions, as shown so well along the Danish coast and at Herschel Island, with shear planes commonly localized in the clay members.

Pore Pressure and Shear Strength

The importance of pore pressure in shear movements within unconsolidated sediments involved in landsliding has long been appreciated. The relief of pore pressure by drainage is, indeed, a standard procedure in stabilizing landslides. Recently, Rubey and Hubbert (21) have indicated the importance of fluid pressure within the pores of a sedimentary succession in facilitating tectonic over-thrusting. Such pore pressure may be equally important in subglacial thrusting.

The critical shear stress (τ_{crit}) necessary for failure or shear of porous sediments and, indeed, for porous fluid-filled solids (10, p. 142) can be expressed by a form of the Mohr-Coulomb law

$$\tau_{crit} = \tau_0 + (S-p)\tan\phi \tag{1}$$

where τ_0 is the cohesion of the sediment, S is the total normal stress, p is the hydrostatic pore pressure, and ϕ the angle of internal friction, like τ_0, a property of the solid.

Soils with a single-grain structure, such as silt, sand, and gravel, with diameters larger than 0.002 mm. are cohesionless so that τ_0 is negligible. Shearing strength is then

$$\tau_{crit} = (S-p)\tan\phi. \tag{1a}$$

The critical shear strength (τ_{crit}) of clays is difficult to evaluate as it is a function of grain properties, water content, and rate of shear. The critical shear strength of saturated clays in quick shear may be dependent only on

τ_0, regardless of applied normal stress, whereas that for partially saturated clays is more complex. The shear strength of mixed soils, where there are relatively small amounts of silt and sand in a true clay, and vice versa, is dependent upon both cohesion and frictional effects, as given in equation (1).

Shear stress applied by a glacier on its substratum has been estimated to be close to 1 kg./cm.^{2} (17; 19; 15). The shear strength of sediments generally exceeds this value, but is lowest in fine-grained and unfrozen soils (16; 24). Abnormal circumstances such as unusually high shear stress beneath the ice, unusually low shear strength of soils, or unusually high pore pressures seem, however, to have permitted shear failure in many widely separated localities.

The Relationships of Total Stress (S_{zz}) and Pore Pressure (p) to Depth

The total stress (S_{zz}) measured on a horizontal surface at depth (z) is determined by the overlying load which is the integral of bulk density (ρ_b), gravitative attraction (g), and depth (dz) from surface to depth z:

$$S_{zz} = \int_0^z \rho_b g dz. \tag{2}$$

In the simple case where density is uniform and independent of depth

$$S_{zz} = \rho_b g \int_0^z dz = \rho_b g z. \quad \text{(see Fig. 1).} \tag{2a}$$

The term bulk density, used in the broad sense followed by Hubbert (7) and Rubey (10) (instead of in the restricted sense followed by Hedberg in which it is applied only to dry sediment) is the weight of both the solid grains and the pore fluid within a unit volume of sediment. If ρ_g = the average grain density, ρ_f = the average fluid density and η the porosity, then for saturated sediments

$$\rho_b = \rho_g(1-\eta)+\rho_f\eta. \tag{3}$$

In the simple case where the sediment is dry and the density of the pore atmosphere is negligible, then

$$\rho_b = \rho_g(1-\eta). \tag{3a}$$

Should such a dry sediment then be saturated with water, the bulk density would be increased by a value determined by the porosity.

The value of the pore pressure (p) is a product of the density of the pore fluid, the acceleration of gravity, and the depth of the zone of fluid saturation. For water it is the depth below the water-table, which can be defined as the level of standing water in megascopic openings (9); this

usage avoids any term for capillary attraction. Where the sediment is saturated to the ground surface and

$$p = \rho_f g z \tag{4}$$

$$\begin{aligned} S_{zz} &= gz[\rho_g(1-\eta)+\rho_f\eta]-\rho_f gz \\ &= gz[\rho_g(1-\eta)-\rho_f(1-\eta)] \end{aligned}$$

and

$$S_{zz}-p = gz(1-\eta)(\rho_g-\rho_f). \tag{5}$$

As grain density in most sediment is in the neighbourhood of 2.65 gm./cm.^{3} and soil water has a density about 1.00 to 1.03 gm./cm.^{3} depending on salinity, the last term in equation (5) has a value of about +1.65 gm./cm.^{3}, the submerged density of the solid particles.

A very different condition prevails within glacial ice.[1] Highly porous firn is limited to a superficial zone, and at depth the porosity (η) approaches zero. Deeper ice may be regarded either as essentially non-porous and impermeable, or as a material in which a limited permeability exists between crystals and along more or less isolated fractures. In the latter case the density of the solid grains, about 0.9 gm./cm.^{3}, is less than that of the interstitial fresh water with density 1.0 gm./cm.^{3}, the last term of equation (5) becomes negative, and $S_{zz}-p$ diminishes rather than increases with depth below the water-table.

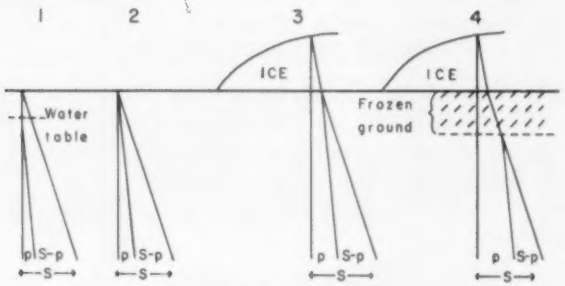

FIGURE 1. Limiting values of effective stress (S-p). Case 1. Water-table beneath the ground surface. Case 2. Water-table at surface. Case 3. Water-table at surface in unfrozen ground beneath a glacier. Case 4. Water-table at base of frozen ground beneath a glacier.

Thus, should a water-table within ice extend to the surface of a glacier and communicate with the glacier floor, ($S_{zz}-p$) at the base of the ice would have a negative value. Were there no adhesion between ice and substratum, the glacier would then be lifted off the ground and movement,

[1]See also in this connection Haefeli, R. (1957). Notes on the formation of ogives as pressure wares; J. Glaciol., vol. 3, no. 21, pp. 27–9.

as "block flow," would be resisted only by the viscosity of a layer of water between them. The generally low surface velocity of glaciers indicates, however, that such a condition could be only temporary or local. In some places on glaciers, especially in areas of compressive flow (18) below the firn line, the water-table lies at or very close to the surface in crevasses, but it is likely that such crevasses do not communicate to depth (i.e., the water-table is perched) and that high uplift pressures need not be assumed for the glacial floor beneath. In other places, particularly in areas of extending flow, open crevasses penetrate to depths of tens of metres and the same value for depth cannot be accepted for ice and for the zone of water saturation. However, high fluid pressures at the base of Salmon Glacier, British Columbia, are suggested (15, p. 453).

Pore Pressures in Unfrozen Ground directly beneath a Glacier

On purely theoretical grounds, high fluid pressures in unfrozen material beneath an *advancing* glacier may be inferred. The addition of a load of ice will tend to bring about compression through closer particle packing and consolidation of a sedimentary substratum by expulsion of pore water. Consolidation will be possible in water-saturated sediments, however, only if water can be expelled. Should there be an unsaturated zone present in the succession before it is covered by the ice, some water can be pressed into this zone with accompanying expulsion or compression of the pre-existing pore atmosphere. Should, however, the sedimentary succession already be completely saturated, consolidation is possible only if there is opportunity for drainage of pore water. If horizontal drainage is to take place entirely within the sedimentary succession, pore pressure at any point must rise according to the rate of expulsion of water, the distance (presumably lengthy) along which the water must move, and the permeability of the sediment. If the ice advance is rapid and the application of ice load fast, high pore pressures may be required to promote significant horizontal expulsion of water and if permeability is low the increase in pressures may approach, though they would not exceed, the increments of total load. If the overlying glacial ice is regarded as impervious, drainage upward is precluded. If the overlying ice is, on the other hand, regarded as slightly permeable but water-saturated, drainage upward is possible but only after fluid pressures within the sediment increase by an amount in excess of that exerted by the fluid column within the ice. If the water-table within the ice lies close to the surface, as at Salmon Glacier, this increase in pore pressure (p) in the underlying sediments may be approximately the same as the increase in total load (S).

The presence within the ice of temperatures well below 0° C. would limit if not preclude the slow movement of water through openings should they extend continuously from surface to base of the ice. Under these circumstances the glacier may well constitute a completely impermeable seal over its substratum.

Effect of Permafrost

The shear strength of soils is usually considerably greater when frozen than unfrozen, because the bonding effect of ice increases τ_0, and the effective normal stress is greater because it is not decreased by hydrostatic pore pressure. Therefore, in a vertical section from the surface of the ground through the permafrost zone to the unfrozen sediments below, the lowest critical shear stress should lie at the base of the permafrost zone (cf. 22). As the overlying sediments, if saturated when frozen, would constitute an impervious layer, drainage upward would be inhibited so that high pore pressures might develop beneath the permafrost zone. If, however, there are clayey beds in the zone of permafrost, shearing may take place in the clay beds, because clay may remain plastic at sub-zero temperatures, even though the clay has included ice segregations and is enclosed above and below by frozen ground. For example, soils along the Yukon Coast with roughly equal portions of clay, silt, and sand fractions were observed to be plastic, even where the soil included ice veins and pellets. In a sedimentary succession with clay members, shearing has often been observed to be confined almost entirely to the clay members, as for example, at Herschel Island and along the Danish Coast.

Frozen ground is almost certainly present beneath the central parts of large polar ice-caps such as those of Greenland (5, 6) and Antarctica (20) where basal ice temperatures of from $-10°$ C. to $-14°$ C. are inferred. It may not be present beneath "temperate" glaciers where surface temperatures are close to 0° C. and, assuming positive temperature gradients, basal ice may be at, but not above, the pressure melting point. Robin (20), however, points out that a negative gradient may prevail in part of a glacier because of movement of cold ice, accumulated at a high level in the névé, downward, and downslope beneath ice accumulated at lower and milder altitudes. Moreover, Mathews (14, p. 96) has noted permafrost exposed by the retreat of a small temperate mountain glacier in southwestern British Columbia.

Although it is not possible to specify, with any precision, the depth of permafrost during periods of glacier advance and retreat, the general factors governing the thickness of permafrost may be estimated.

Thus if:

T_l = the mean annual temperature in °C. for the land surface or ocean bottom;

T_i = the mean annual basal ice temperature in °C.;

i_g = geothermal gradient in metres per °C.;

Z_l = depth in metres to base of permafrost under the land surface under steady state conditions;

Z_i = depth in metres to base of premafrost beneath a glacier under steady state conditions;

Z_{lst} = depth to base of permafrost, after (t) years of ice cover at point

(x), where the glacier front has advanced a sufficient distance so that the development of permafrost is now governed by the thermal régime of the glacier. For practical purposes, an advance of 1,000 metres in ten years would be adequate in distance and time.

Under steady state conditions, and assuming a linear variation of temperature with depth, the thickness of permafrost (12, p. 1522; 23, p. 27) in the ground in metres is approximated by

$$Z = T_l/i_g. \tag{6}$$

If a glacier advanced over the land or ocean bottom past position x, permafrost conditions at time t would be approximately as follows:

(a) If $T_l > 0$, probably a rare circumstance except near the southern margin of glaciation, there would be no initial permafrost. Over a period of years, permafrost would begin to grow downwards beneath the glacier, the rapidity depending upon such factors as the ground temperature (T_l) before the ice advance, material, porosity, and T_i. In Figure 2, the approximate growth of permafrost under glacier ice at $T_i = -2.5°$ C. and $-5.0°$ C. with T_l at 0° C. is plotted, assuming saturated but unfrozen ground with a porosity of 25 per cent, .2 cal./gm./°C. average heat capacity of the soil particles, and 6×10^{-3} cal./cm./sec./°C. the average thermal conductivity of the soil (23, pp. 39–43). Latent heat of fusion of ice is considered in the calculations. Starting with ground temperatures (T_l) of 0° C. Figure 2 shows that permafrost tens of metres in thickness might develop within a few score years.

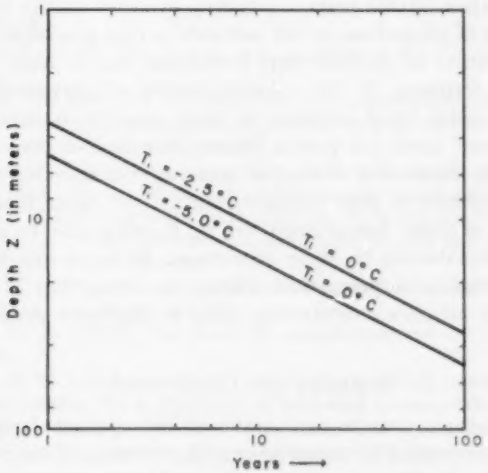

FIGURE 2. Graph of growth of permafrost where T_i is the mean annual basal ice temperature and T_l the mean annual ground temperature.

If the ice advanced over an unfrozen lake bottom 1,000 or more metres in diameter, with water depths exceeding the maximum winter thickness of ice, the mean annual temperature of the lake bottom would be above 0° C. In that case, permafrost would begin to form from the bottom of the lake downwards. If the ice advanced over the sea floor, ocean-bottom temperatures might be above or below 0° C. At Barrow, Alaska, the mean annual ocean-bottom temperature is estimated to be −0.5° C. and permafrost 20–40 metres in thickness may extend seaward (12, p. 1523). However, as the pore water in the ground is saline, ice may not be present in the permafrost zone.

(b) If $T_i < T_l < 0$, then $Z_i > Z_l > 0$ and Z_{lzt} will increase with time to approach Z_i.

(c) If $T_l < T_i < 0$, then $Z_l > Z_i > 0$ and Z_{lzt} will decrease in thickness. The decrease in thickness of permafrost is relatively slow. Terzaghi (23, pp. 30–1) has shown that in frozen saturated soils the upward rate of permafrost recession due to heat radiation from the interior of the earth is unlikely to exceed two centimetres per year and is probably less. Therefore, it would take many thousands of years to reduce an area of thick permafrost, say 400 metres, to a thickness of 100 metres, which represents the near maximum thickness of sheared beds. Consequently, if saturated frozen unconsolidated sediments were in thermal equilibrium with mean annual ground temperatures comparable to those, for example, of modern temperatures in the Mackenzie delta area (about −10° C.) and the area were overridden by a glacier, permafrost 100 or 200 metres thick should be present even after 10,000 years.

The growth of permafrost by the advance of cold glacial ice over warmer land, lake bottom, or sea floor may contribute also to high pore pressure, and hence to thrusting, by the volume increase of interstitial water in the process of freezing. Such increase in pore pressures is noted in seasonal freezing in many areas (2, p. 6). During recession of permafrost, on the other hand, the decrease of volume of pore material from ice to water could lead to a reduction of pore pressure, but on the other hand the loss in cohesion (τ_0) of many fine-grained soils on thawing may be a more important factor contributing to shear movement. Perhaps also favourable for shear is a recession of permafrost leading to weakening of the sediment followed by an advance contributing again to high pore pressure.

Summary and Conclusions

Glacially deformed sediments are typically arcuate in plan and are believed to have formed by thrusting near the terminus of an actively moving lobe of ice. The ice-thrust features range in thickness up to a maximum of some 200 metres. Available evidence commonly suggests that mixed and cohesionless soils, if not the clays, were frozen when deformed, thus indicating the presence of permafrost. As the permafrost zone would provide an

impervious barrier to the escape of pore water upward, high pore water pressures might have developed beneath it. Whenever pore pressures increased towards the limiting value of the normal stress, the effective normal stress, and hence τ_{crit}, would have approached the value of τ_0 (Figure 1). Once movement started, τ_0 would have been virtually eliminated and further movement would have been greatly facilitated.

If this is correct, then τ_{crit} would have reached a minimum value at the base of the permafrost layer, except in some successions containing plastic clay members, and the thickness of the permafrost zone would govern whether shearing would or would not occur. Permafrost of moderate thickness would probably have been present: (1) where ice advanced over a land area with mean annual ground temperatures a few degrees below 0° C.; (2) where a glacier with basal ice temperatures several degrees below 0° C. advanced over an unfrozen lake or sea floor close to 0° C.; or (3) where permafrost of limited thickness was present beneath a glacier in which after a relatively brief period of retreat the ice readvanced. The preceding three conditions are believed to be sufficient for a moderate thickness of permafrost to occur under an advancing ice lobe and for pore pressures to build up.

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SECTION FOUR

A Kinderhook Microfauna from Crowsnest Pass, Alberta*

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Presented by J. F. CALEY, F.R.S.C.

ABSTRACT

A previously unrecorded microfauna consisting primarily of ostracods and conodonts has been obtained from strata of the Exshaw formation at Crowsnest Pass, Alberta. It contains several Kinderhook species previously described only from the mid-continental United States. An early Mississippian age is indicated for this occurrence.

THE position of the Devonian-Mississippian boundary in the Rocky Mountain region of Western Canada has long been a matter of discussion (*see* 18). It is generally assumed to occur between the Palliser and Banff formations within the stratigraphic interval occupied by the Exshaw formation. Recently Harker and McLaren (9) have suggested that the well-defined contact of the Exshaw and underlying Palliser formations may possibly mark this boundary. Faunally, the Palliser formation, in its upper part, bears the Upper Devonian *Leiorhynchus utahense ventricosum* fauna and the Banff carbonate contains several Lower Mississippian (Kinderhook) species (9, p. 252). Between these two stratigraphic units lies a thin sequence of predominantly black shale beds containing at a few localities a sparse brachiopod (9) and cephalopod fauna (19). These strata are broadly divisible into three units: the lower black shale (occasionally with a basal sandstone), and middle siltstone (locally a silty limestone) units comprising the Exshaw formation; and the upper black shale, the basal unit of the Banff formation. Palaeoecological conditions of deposition of these beds were probably somewhat similar to those proposed by Gutschick and Perry (8, p. 979) for the Sappington formation of Montana.

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In the Crowsnest Pass (sec. 8, tp. 8, rge. 5, W. 5, Alberta) on the southern side of Alberta–British Columbia Highway No. 3 along Crowsnest Lake, the sequence of beds in the stratigraphic interval between the top of the Palliser formation and the base of the carbonate portion of the Banff formation has an aggregate thickness of approximately fifty-four feet (Fig. 1). Slight structural disturbance of minor importance is evident within the upper shale unit and, at the Palliser contact, an unknown thickness of the basal black shale and uppermost beds of the Palliser formation may be missing due to faulting. This sequence is comprised of three stratigraphic units: a basal sandy black shale, an intermediate silty limestone with intercalated grey shale beds, and an upper sandy black shale. These units are interpreted respectively as representing the lower and upper portions of the Exshaw formation and the basal unit of the Banff formation. This interpretation appears to be consistent with the lithological sequence exemplified by the Exshaw type section on Jura Creek. A somewhat different evaluation of the stratigraphic relationships of these units at Crowsnest Pass has been advanced by de Wit (6, p. 106, photograph 9, p. 229). He considers the presence of glauconite in the two upper units as a possible indication of their relation to the Lodgepole formation of Montana and includes both in the Banff formation.

Collections for microfossil analyses were taken systematically across this sequence except for two covered intervals of six and seven feet respectively within the lower shale. Samples were also collected from a faulted five-foot bed of unfossiliferous black shale about forty feet above the base of the Banff carbonate. The shales were prepared for examination by the conventional method of soaking in water; the silty limestone samples were disintegrated in weak formic acid. All specimens appear to be etched either by weathering processes or other chemical action, they are not abraded and thus are not thought to have been transported prior to burial.

Fauna was obtained only from the interval between twenty-three and forty-two feet above the top of the Palliser limestone. It consisted of crinoid columnals, rare brachiopod and cephalopod (?) fragments, a generalized zaphrentid coral, fifteen species of nine genera of ostracods, at least four genera of conodonts, and one species questionably referred to the Foraminifera.

Only one occurrence of ostracods has been mentioned previously from this stratigraphic interval in the Canadian Rocky Mountain region (18, p. 965; 19, p. 971). Gutschick and Perry (7, p. 1896; 8, p. 980) mention the presence of ostracods, and Knechtel and Hass (12), Cooper and Sloss (5), and Achauer (1) have recorded the occurrence of conodonts from the "Sappington" formation and from a black shale—subsequently named the Little Chief Canyon—member of the Lodgepole formation (13) in Montana. Their work indicates a Lower Mississippian (Kinderhook) age for these beds. Possible correlation of the fossiliferous strata of the Crowsnest Pass section with the basal black shale of the Lodgepole forma-

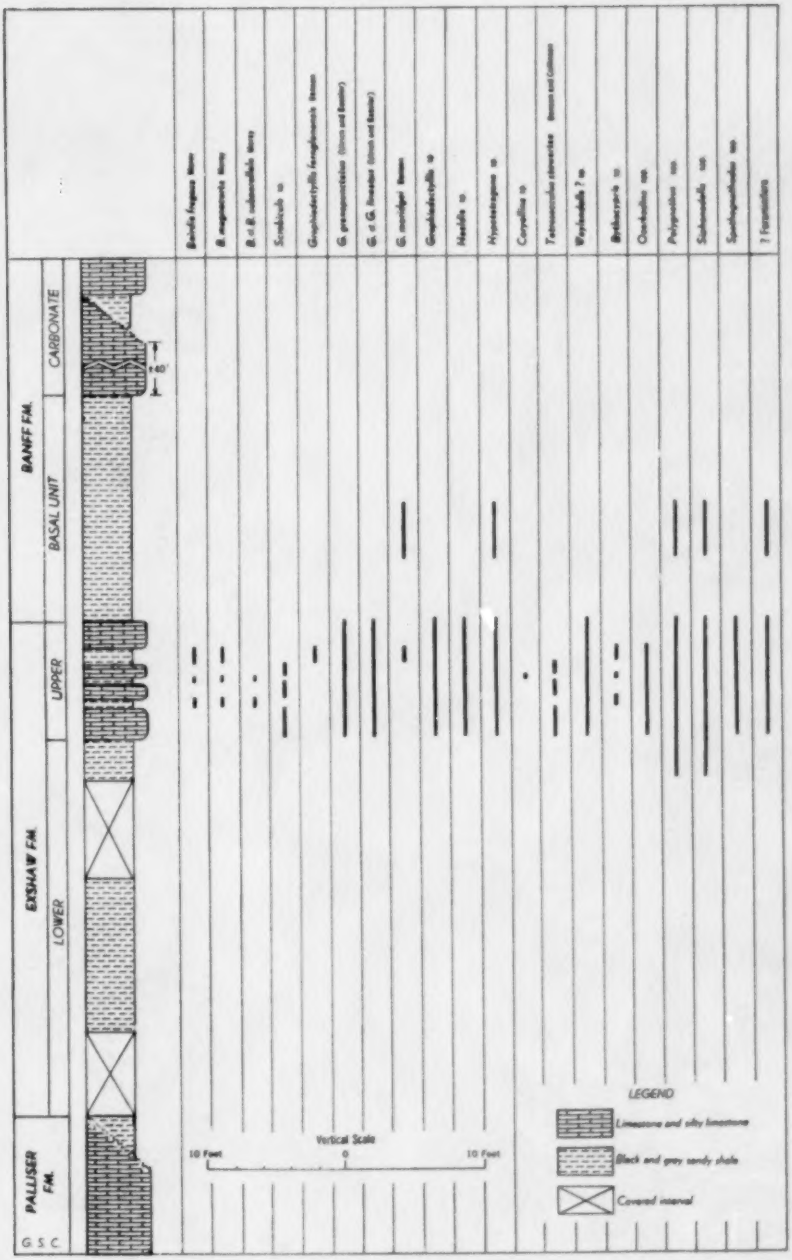

FIGURE 1. Stratigraphic occurrence of microfauna, Crowsnest Pass, Alberta.

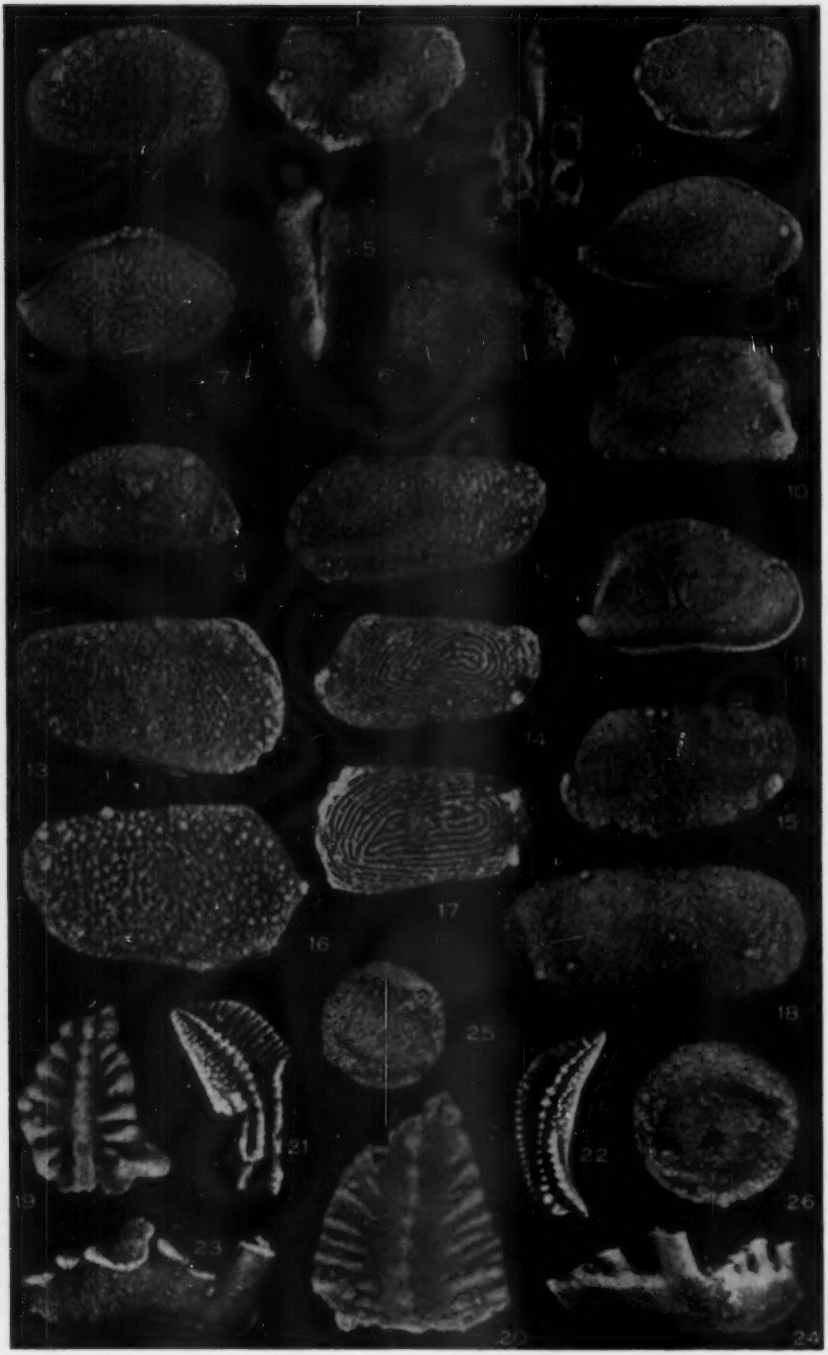

PLATE I

tion in Montana may tentatively be based on the common occurrence of the distinctive conodont *Siphonodella* (= *Siphonognathus* Branson and Mehl, 1934 [not Richardson, 1858]) *duplicata* (Branson and Mehl) (*see* 12, p. 519). Correlation may also be possible with the "Sappington" formation of southwestern Montana based in part on the stratigraphic occurrence of similar conodont genera. Achauer (1) described the "Sappington" of that area as divisible into three units—basal and upper black shale members bearing sparse conodont assemblages and a middle argillaceous and calcareous shale and siltstone member containing, in its upper part, a conodont assemblage prolific in specimens of the genera *Siphonodella* and *Polygnathus*. This lithology and conodont distribution is repeated in the Crowsnest Pass sequence where the major conodont assemblage (consisting predominantly of specimens of *Siphonodella* and *Polygnathus*) is found in the middle silty limestone and intercalated grey shale unit. A somewhat similar threefold sequence of conodont-bearing beds comprises the subsurface Bakken formation (Exshaw equivalent?) of southern Saskatchewan and Manitoba. These strata have been described by MacDonald (15), Kents (11), and McCabe (14), but their conodont faunas are as yet unknown. Little apparent similarity is evident, however, between the Crowsnest Pass conodonts and those described from Montana and central Alberta by Cooper and Sloss (5).

Correlation with the mid-continent region of the United States is uncertain. Holland (10, p. 1709) stated that ". . . the ["Sappington"] fauna shows close affinities with the fauna of the Louisiana limestone of Missouri."

PLATE I

(All figures approximately × 32)

1. *Scrobicula* sp. Right lateral view of a complete carapace; hypotype, GSC no. 15165. 2, 3. *Tetrasacculus stewartae* Benson and Collinson. Left lateral and ventral views of a complete carapace; hypotype, GSC no. 15157. 4. *Coryellina* sp. Left lateral view of an incomplete carapace; hypotype, GSC no. 15166. 5, 6. *Hypotetragona* sp. Dorsal and right lateral views of a carapace; hypotype, GSC no. 15161. 7. *Bairdia magnacurta* Morey. Right lateral view of a carapace; hypotype, GSC no. 15168. 8. *Bairdia fragosa* Morey. Right lateral view of a carapace; hypotype, GSC no. 15159. 9. *Bairdia* cf. *B. subparallela* Morey. Left lateral view of an incomplete carapace; hypotype, GSC no. 15167. 10, 11. *Healdia* sp. Left and right lateral views of two carapaces; hypotypes, GSC nos. 15158, a. 12, 13. *Graphiadactyllis fernglenensis* Benson. Left and right lateral views of two carapaces; hypotypes, GSC nos. 15160, a. 14. *Graphiadactyllis* cf. *G. lineatus* (Ulrich and Bassler). Left lateral view of a complete carapace; hypotype, GSC no. 15155. 15. *Graphiadactyllis moridgei* Benson. Left lateral view of a complete carapace; hypotype, GSC no. 15163. 16. *Graphiadactyllis granopunctatus* (Ulrich and Bassler). Right lateral view of a complete carapace; hypotype, GSC no. 15162. 17. *Graphiadactyllis* sp. Left lateral view of an incomplete carapace; hypotype, GSC no. 15156. 18. *Waylandella* ? sp. Right lateral view of a complete carapace; hypotype, GSC no. 15164. 19, 20. *Polygnathus* spp. Two incomplete specimens; hypotypes, GSC nos. 15171, a. 21, 22. *Siphonodella duplicata* (Branson and Mehl). Two complete specimens; hypotypes, GSC nos. 15170, a. 23, 24. *Ozarkodina* sp. Two broken specimens; hypotypes, GSC nos. 15172, a. 25, 26. ? Foraminifera sp. Views of two specimens; hypotypes, GSC nos. 15173, a.

This cannot be shown on the basis of similar microfauna as the micropalaeontological content of these formations has not been extensively studied. Ostracod faunas from Illinois (2; 3) and Missouri (16; 17) bear, in general, a resemblance to the present fauna. The species identified from the Crowsnest Pass locality (Table I; Pl. I) are similar to species described from the Kinderhook (Bushberg, Chouteau, Springville, etc.) and Lower Osage (Fern Glen) series of the mid-continent region (20; 3).

TABLE I

STRATIGRAPHIC OCCURRENCE OF CROWSNEST PASS MISSISSIPPIAN OSTRACOD SPECIES IN MID-CONTINENTAL NORTH AMERICA (AFTER BENSON AND COLLINSON (3), TABLE I, IN PART)

Crowsnest Pass species	Fern Glen	Springville	Chouteau	Bushberg
Graphiadactyllis cf. *G. lineatus*	X	X		
Graphiadactyllis fernglenensis	X	X		
Graphiadactyllis granopunctatus	X	X	X	X
Graphiadactyllis moridgei	X	X	X	
Tetrasacculus stewartae		X	X	
Bairdia fragosa				?
Bairdia cf. *B. subparallela*				?
Bairdia magnacurta	X	X	X	X
Hypotetragona sp.				?

The Crowsnest Pass fauna, therefore, indicates for the first time the extension of these Lower Mississippian ostracods to the Rocky Mountain region. Based on the asserted Kinderhook age of the megafauna from the middle part of the overlying Banff carbonate sequence (9, p. 252) an earlier Kinderhook age is assume for the Exshaw microfauna-bearing beds at Crowsnest Pass. This agrees in part with Schindewolf (19, p. 975) who, on the basis of the Exshaw cephalopod fauna contained in an undifferentiated section near the Brazeau River, Alberta, stated: "The Exshaw formation is thus definitely of Lower Mississippian age and probably equivalent to the Chouteau formation of Missouri (Tournaisian of European terminology)." Whether this probable equivalence of the Exshaw to the Chouteau can be proved for the entire formation or whether the Exshaw may be equivalent to more than one mid-continental formation of Kinderhook age (as is quite possible) is not known. Unfortunately, the only partially exposed twenty-three feet of basal sandy black shale at the Crowsnest Pass locality has as yet failed to yield a microfauna which might indicate the equivalence of these beds and more exactly define the base of the Mississippian in this area.

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TRANSACTIONS OF THE ROYAL SOCIETY OF CANADA
VOLUME LIV : SERIES III : JUNE, 1960
SECTION FOUR

Jurassic-Cretaceous Boundary in Western Canada and Late Jurassic Age of the Kootenay Formation

WM. C. GUSSOW, F.R.S.C.

ABSTRACT

Regional structural relations together with faunal and floral evidence demonstrate that the Jurassic-Cretaceous boundary can now be drawn with some confidence in Western Canada. The basal Blairmore conglomerate of the Rockies and Foothills and the basal quartzose sandstone of the Plains mark the base of the Cretaceous, and are separated by a major regional unconformity from the underlying eroded strata which range in age from late Jurassic to Mississippian and Devonian or older. The Kootenay formation is assigned to the Portlandian-Purbeckian stages of Europe, and the basal Blairmore to the Wealden 4 (Berriasian). The Blairmore represents the entire Lower Cretaceous, ranging in age from Wealden 4 to 6, through Valanginian, Hauterivian, Barremian, Aptian, and Albian. The Canadian evidence strongly supports the boundary suggested recently for the European standard section.

MANY great geologists of the past and present have wrestled with the problem of the age of the Kootenay formation in Western Canada. This is the source of important coal deposits and, accordingly, there has been much study of its contained flora. About a hundred years ago, James Hector (34) made the first geologist's report on the presence of important coal deposits in Western Canada. Seventy-five years ago, in 1885, Sir William Dawson (19, pp. 2–10) presented a paper before this Society and described the Mesozoic floras of the Rocky Mountain region of Canada. He considered this flora to be early Cretaceous. This material was collected and submitted by G. M. Dawson who made the first geological map and report on the Rocky Mountains and Foothills area south from the Kicking Horse Pass to the Forty-ninth Parallel (18). In this report, Dawson introduced the term "Kootanie series," which included the present Fernie and Blairmore. W. W. Leach (43) differentiated the Fernie and Kootenay formations, but named the basal Blairmore conglomerate the Kootenay conglomerate. In 1915–16, B. Rose (66, pp. 109–10) recognized the unconformity at the base of the conglomerate and accordingly placed it in its proper position at the base of the Blairmore formation. Lammers (39) correlated the basal Blairmore conglomerate with the Cloverly conglomerate, which he considered to mark the base of the Cretaceous. Accordingly, he assigned an Upper Jurassic age to the Canadian Kootenay.

"The boundary of the Cretaceous system is placed very precisely where marine strata assigned by general agreement to the earliest part of the Berriasian stage rest upon marine strata similarly assigned to the latest part of the Tithonian stage, as in the western Alps. In many regions, however, no such precision is possible, for some of the beds may be nonmarine or unfossiliferous or otherwise difficultly assignable, and opinions may differ for each region concerning the position of the boundary." (15, pp. 1012–13)

Geologists have long recognized a hiatus in the fossil record of the Fernie-Blairmore succession in Western Canada. Until recently, the youngest known marine fauna from the Fernie ranged up into the Oxfordian; the upper flora of the Blairmore had been assigned to the Albian with some confidence (9, p. 11); and the lower flora of the Blairmore was assigned an age not much older—the Aptian. A diagnostic record of the Jurassic Kimmeridgian, Portlandian, Purbeckian, and the Cretaceous Neocomian appeared to be missing. On the basis of the contained flora, Bell (9, p. 8) had always favoured a Neocomian-Barremian age for the intervening Kootenay formation. In discussing the age of the Kootenay, Dannenberg (17, 266) noted that the over-all similarity to Jurassic forms is most striking and the absence of dicotyledons in the Kootenay is disturbing.

As pointed out by McLearn (56, p. 157) nearly thirty years ago, "There is a time in the history of every study when it is advisable to pause, appraise what has been done, determine important gaps in the evidence, and plan how and where these gaps may be filled." This time has come again in the investigation of the Jurassic-Cretaceous problem of the Canadian interior. In recent years, much new information has come to hand as a result of the great activity in exploration for oil and gas and, furthermore, much new information has been published with regard to the European standard section. The establishment of a reliable chronology is fundamental and cannot be overemphasized.

Ostracods have played an important part in working out the true Jurassic-Cretaceous geological boundary in Europe. This problem was studied recently by Bartenstein and Burri (4). This geological bounary can be demonstrated even more dramatically in Western Canada, on a structural basis, and can be supported on palaeontological grounds.

In 1955, Gussow (33, p. 143) discussed the desirability of placing the Jurassic-Cretaceous boundary at the base of the Cadomin or Blairmore conglomerate rather than within the underlying conformable Kootenay-Fernie sequence. Frebold (30, p. 143) agreed: "It would certainly be easier and much more convenient for the field geologist to place the top of the Jurassic at the Cadomin or Blairmore conglomerate. It would, however, be necessary to prove a Jurassic age for all of the Kootenay formation and its equivalents. Presently we only know that the lowermost Kootenay sandstone is Jurassic in age, i.e. upper Portlandian, as proved by the presence of *Titanites*."

Stratigraphic Relations

In sections along the Foothills belt, from southeast to northwest, the most persistent lithological unit is the conglomerate at the base of the Blairmore formation (74). This forms an excellent mapping unit that is now recognized by all geologists working in the disturbed belt. Continuing to the northwest, this conglomerate is referred to as the Cadomin formation (49) in the central Foothills, and is the upper conglomeratic part of the "Dunlevy" in the Peace River vicinity. To the south in Montana and Wyoming, it is referred to as the Cloverly conglomerate (Pryor conglomerate, Lakota sandstone, 3rd Cat Creek sand, or Sunburst sand). All are now known to be the same unit. Above this unit lie the Blairmore, the Luscar, the Gething, and the emended Kootenai (14, 11) of Montana, which are in part equivalents, and below the basal conglomerate unit are the Kootenay and Nikanassin, and the lower "Dunlevy," which are contemporaneous.

"In the foothills of southern Alberta . . . the Blairmore group is readily differentiated lithologically from the underlying Kootenay formation by the almost complete lack of coal, and by a basal chert-pebble conglomerate which in places lies upon the Kootenay with a marked erosional unconformity." (9, p. 2) East of the Foothills belt, the basal Blairmore unconformity is clearly defined as the Kootenay formation has been entirely removed by erosion and the limiting beds at the unconformity become farther apart in age, so that basal Blairmore rests unconformably on marine Jurassic, Mississippian, Devonian, and older strata, eastward (Fig. 1). In the Foothills, the basal Blairmore has now been defined by regional mapping, as lying unconformably on the Kootenay and/or Nikanassin. To the west, however, in the Fernie Basin, considerable difficulty arises as the Kootenay becomes more sandy and conglomeratic, with the result that some geologists (66; 51, p. 5; 1, pp. 28–9) have picked an arbitrary boundary within the Kootenay and placed the conglomeratic member in the Blairmore.

The Kootenay and Nikanassin both belong to the Jurassic sequence and are conformable and transitional with the underlying Fernie (12; 6; 20, p. 22; 23; 36, p. 16; 16, p. 66; 24, p. 149; 72; 61; 62). Rose (66, p. 109) described the base of the Kootenay as a gradual change from marine shales to subaerial sandstones, and Stewart (71, p. 25) arbitrarily defined the base of the Kootenay as a greenish thin-bedded sandstone, seventy-five feet thick, conformably underlain by the Fernie, and stated that the passage from the Fernie to the Kootenay is transitional.

The type section of the Kootenay formation was recently proposed by Norris (62) as the south face of Grassy Mountain, five miles north of Blairmore, Alberta. The section was divided into four units: from the bottom up, the Moose Mountain (named by Beach [5]), Adanac, Hillcrest, and Mutz members. A fifth member, the Elk formation, is found to the west and is here considered to belong to the Kootenay. At the type locality, the lower

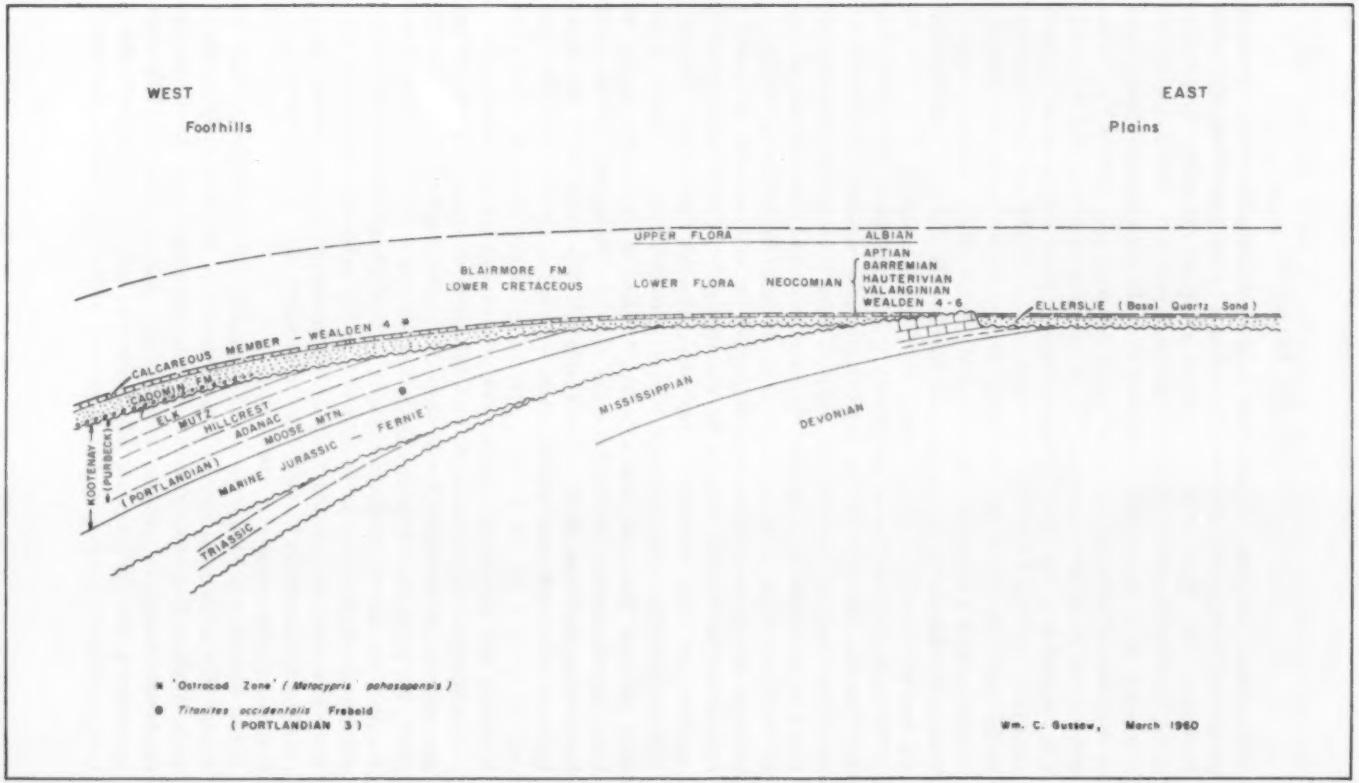

FIGURE 1. Diagrammatic section showing Cretaceous-Jurassic relations in Western Canada.

TABLE I
SIMPLIFIED CORRELATION TABLE

		Alberta and British Columbia		
Wyoming	Montana	Southern Foothills	Central Foothills	Peace River area
Cloverly fm.	Kootenai fm.	Blairmore fm.	Luscar fm.	Gething fm.
Cloverly congl.	Kootenai congl.	Blairmore congl.	Cadomin fm.	upper "Dunlevy" (congl.)
		REGIONAL UNCONFORMITY		
	Enlarged	Kootenay fm.	Nikanassin fm.	lower "Dunlevy" (sh & ss)
	Morrison	Passage beds	Passage beds	
	Ellis gp.	Fernie fm.	Fernie fm.	

contact with the Fernie is described as gradational, and the upper contact as a regional unconformity, but locally beds above and below this are concordant. (See Fig. 2, col. 4.)

THE KOOTENAY-BLAIRMORE AGE CONTROVERSY

In general, two floras are present in the Blairmore formation, and the Canadian Kootenay carries a third. One of the most characteristic species

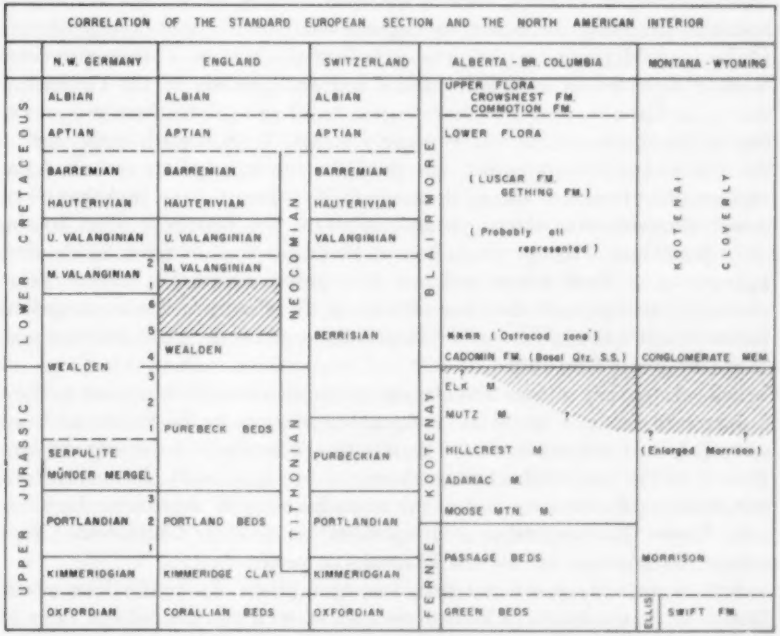

FIGURE 2. Generalized geological correlation of the Upper Jurassic and Lower Cretaceous. Columns 1, 2, and 3, in part after Bartenstein and Burri (4).

of the Kootenay is *Cladophlebis heterophylla*. As the plants found in the Kootenay formation are long-ranging types, the exact age of the Kootenay has always been a matter of controversy (52, 12, 21, 54, 55, 78, 56, 71, 10, 73, 11, 8, 9). In 1914, D. D. Cairnes (12, p. 32) observed:

> The fossils, chiefly plants, found in this formation [Kootenay] in different localities, point very strongly to the possibility of its being Jurassic and not Cretaceous, as has been previously supposed. The Fernie shales, which lie immediately beneath, have been proved by Mr. T. W. Stanton, and others, to be Jurassic; and in this region in the foothills no evidence of an unconformity or any lapse of time exists between the Kootenay and the Fernie shales (in fact they gradually change into one another); and it would seem possible that the conglomerate at the top of the Kootenay might mark a slight time hiatus, particularly as the rocks above and below are so entirely different in character—both in colour and in character of their components. So for these reasons it is considered very probable that the coal measures, as well as the Fernie shales, are Jurassic, and that the Dakota, in this region, forms the bottom of the Cretaceous.

In 1915, McLearn (54, p. 112) placed the Kootenay in the Upper Jurassic because of the apparent conformity of the Kootenay and the Fernie and the unconformity between the Kootenay and the Blairmore.

A special tribute is due to Walter A. Bell, palaeobotanist with the Geological Survey of Canada, who in the short span of forty years has single-handedly prepared, identified, catalogued, compiled, figured, and published all our knowledge on the known fossil flora of all Canada. This monumental work is recorded in various memoirs and publications of the Geological Survey of Canada. Bell must have encountered great difficulties in pioneering the fossil plant studies in Western Canada. With regard to the age of the Canadian Kootenay, Bell (8, p. 513) correctly points out that the stratigraphic ranges of species do not indicate a Jurassic age, and that more genera characteristic of the Jurassic occur in the Blairmore than in the older Kootenay. The age of the lowest Blairmore flora is fixed by the first appearance of dicotyledons and was accordingly assigned a Lower Cretaceous Aptian age, and the close affinity of the Kootenay floras compelled Bell to assign a Barremian or late Neocomian age for the plant-bearing part of the Kootenay formation.

Bell (8, p. 517) admits that the age of the Kootenay formation could be Upper Jurassic, in so far as its stratigraphic relations to the Fernie are concerned. As he considers the lower flora of the Blairmore to be wholly Aptian, Bell (8, p. 522) concludes that as the upper age limit of the Kootenay flora is Aptian (or Barremian) and as the lower limit set by the marine Fernie is early Upper Jurassic, either a late Jurassic or an early Cretaceous (Neocomian-Barremian) age for the Kootenay is possible.

Bell (8, p. 522) points out that a revision of the Kootenay flora in the Blairmore area reduces the known species to fifteen and that all but three of these were collected from beds lying within the upper half of the Kootenay. The first and most notable feature of this flora is its almost entire lack of

individuality, for of the fifteen species only three are confined to the Kootenay—the balance all occur in the Blairmore, or Luscar and Gething equivalents. This range of species certainly does not suggest the presence of any appreciable time-break at the Kootenay-Cretaceous boundary. The three species confined to the Kootenay are *Sphenopteris* (*Adiantum* ?) *montanense* (Know.), *Nilssonia schaumburgensis* (Dunk.), and *Czekanowskia dichotoma* (Heer). *N. schaumbergensis* resembles forms in the Jurassic but is also known in beds younger than Kootenay. As pointed out by Bell (8, p. 523), *Czekanowski dichotoma* strongly suggests a Jurassic age, even the genus is essentially Jurassic, and *C. dichotoma* is apparently closely allied to, if not actually conspecific with, the Jurassic *C. rigida*. Furthermore, Bell states that another Jurassic holdover, *Podozamites lanceolatus*, occurring in the Kootenay, ranges upward into the Blairmore and equivalent. Bell also points out that *Ctenis albertensis* Warren (= *Ctensis borealis* Dawson), claimed by Warren (73) to indicate a possible Jurassic age, occurs also, though rarely, in the Luscar formation. Warren (73, p. 49) raised the question "whether the age of the Kootenay flora has not been placed too high."

With reference to the Nikanassin, Bell (9, p. 20) states that very little is known as yet about its flora, and that all plants of known stratigraphic position were collected from near the top of the formation (actually, the eroded top). Among those listed are *Czekanowskia* cf. *rigida*, *Podozamites lanceolatus*, and *Nilssonia nigracollensis*, which have strong Jurassic affinities and except for the long-ranging *P. lanceolatus*, are not known to occur above the unconformity.

According to McLearn (58, p. 10) a small species of *Aucella* from basal sandstones of the "Dunlevy" formation (now obsolete and recognized as Nikanassin) and from the Nikanassin, although indeterminate, suggest small Jurassic species. Lang (41) also reports that fossils found in the lower Nikanassin of the Moberly Creek area are Jurassic in age. Lang (40, p. 12) notes that the Nikanassin is over 1000 feet thick in the Brulé map-area and may be marine and Jurassic in age in its lower part. In the Hay Mountain region, Spivak (69, p. 539) notes that the lower 750 feet of the Nikanassin is probably marine and contains *Aucella* which may be Upper Jurassic in age. In the Cabin Creek area, pelecypods are found in the upper part of the Nikanassin and are tentatively identified as Upper Jurassic (69, p. 540) and a large part of the section is probably marine in origin. In the Copton Creek area, the Jurassic is overlain by more than 2000 feet of Nikanassin sandstone and shale without apparent break and contains Jurassic marine fossils (Tithonian) in the lower 1600 feet (69, p. 540).

In 1949, a student helper of Jack Crabb on a survey party under Newmarch (60; J. A. Noble, personal communication) made a discovery of great significance. This was a gigantic ammonite at the top of the lowermost Kootenay sandstone which is now referred to as the Moose Mountain member. In 1949, Jeletzky (*in* 60, p. 47) tentatively assigned the ammonite

to uppermost Upper Jurassic rather than Lower Cretaceous. This was later identified and described as *Titanites occidentalis* Freb. n.sp. (29, p. 35) and assigned to the Upper Portlandian (Portlandian 3) of the European standard section. This find is of considerable importance as it determines the age of the Moose Mountain member, and fixes the age of the lower Kootenay formation, and restricts the underlying Passage beds to not younger than Lower Portlandian.

Although Frebold (29,p. 35) still places the bulk of the Kootenay in the Lower Cretaceous, he did mention the possibility that the lower part, above the basal sandstone, could be of Purbeck age. Lord (48, p. 11) reports that Frebold studied "the upper part of the Jurassic system in the central foothills of Alberta" and that "this work suggested, but has not yet proven, that the Nikanassin strata are at least partly of late Jurassic age, and that the remainder are of Cretaceous age."

No distinctive index fossils have been found in the Upper Jurassic (Passage) beds which lie above the Lower Oxfordian Green beds (29, p. 32) and below the basal Kootenay standstone.[1] However, as the so-called "Passage beds" occur below the Upper Portlandian Moose Mountain member and as there is no evidence of any hiatus in the sequence, they may be assigned to the Upper Oxfordian, Kimmeridgian, and Lower Portlandian stages.

Rouse (67, p. 303), in a paper on the "Lower Cretaceous" microflora from the Kootenay formation of southeastern British Columbia, states that the comparison with microfloras from Europe and the U.S.S.R. suggests a closer relationship of the Kootenay to the Upper Jurassic than to the Lower Cretaceous. This conclusion is based mainly on similarity among spores and the absence of striated spores in the Kootenay. The latter are numerous in kind and number in microfloras of the Wealden (presumably, Cretaceous Wealden 4 to 6 ?) and the Lower Cretaceous of the U.S.S.R. (67, p. 305).

In work based on microfloral assemblages of spores and pollen, Pocock (64) has demonstrated that those collected from the Basal Quartz Sand (Ellerslie) and from the youngest recognized beds of Jurassic age are quite distinct; the former bearing a strong resemblance to assemblages from known Lower Cretaceous beds, while the latter have obvious affinities with underlying Jurassic strata. Pocock concludes, on the basis of the general similarity of histograms, that deposition of the Quartz Sand series was more or less contemporaneous over the whole of Western Canada, and that they can be assigned to one definite stratigraphic position. Similarly, McLearn (58, p. 9) was able to demonstrate by the use of non-marine faunas, that the lower part of the Blairmore in the Blairmore area—where it includes *Unio*

[1]*Buchia concentrica* (Sowerby) occurs nine feet below the base of the Passage beds in the southern Canadian Rockies and Foothills (31, p. 3). In the Miette area east of Jasper, *Buchia* was found sixty to ninety feet above thin glauconitic (Green) beds (31, p. 7) in strata assigned to the Passage beds, but which probably correlate with the uppermost Green beds to the south.

(*Quadrula*) *natosini*, *Unio* (*Elliptio*) *hamili*, *Unio* (*Elliptio*) *douglassi*, *Sphaerium onestae*, and *Corbula? palliseri*—is coeval with the lower part of the Blairmore in the Pekisko Forest Reserve farther north, and on the Elbow River, and in the Plains in a well near Lethbridge. It is also equivalent to the Luscar in the central Foothills, to the Kootenai of Montana, and to the McMurray (tentatively) of northern Alberta.

Pocock also draws attention to the striking and abrupt change in character of microfloral assemblages at the Purbeck-Wealden junction, which indicates a previously unsuspected floral break at that point and provides strong palaeontological evidence for placing the Jurassic-Cretaceous boundary in this position. He then demonstrated the striking resemblance of the microfloras of the Purbeck and Wealden (of England) to the youngest Jurassic and Basal Quartz Sand series of Western Canada. Pocock also states (personal communication) that pollen studies demonstrate that there is a complete gradation in evolution throughout the Blairmore, so that all stages are undoubtedly represented.

The upper flora of the Blairmore was at first dated as Cenomanian (early Upper Cretaceous) by Berry (10, p. 56), but it was pointed out by McLearn (78, p. 103; 56, p. 161) that it is at least as early as Albian (late Lower Cretaceous) because the marine fauna above it is late Albian (70, p. 408). In McLearn's opinion, "the dating of the lower Blairmore flora on plant evidence alone is not satisfactory" (56, p. 162). However, he considered it as probably Aptian in age. The lower part of the Blairmore formation was known to contain a non-marine invertebrate fauna, the *Unio hamili* fauna, and in northeastern British Columbia, in the foothills of the Peace and Pine rivers, the lower flora of the Blairmore lies below a *Gastroplites* fauna (early Albian) and the overlying *Neogastroplites* fauna (late Albian) and is thus pre-Albian, i.e. at least as old as Aptian (56, p. 166). A total of five marine and partly marine faunas were recognized in the Blairmore of the Plains and Foothills. These are (56, p. 167):

5.	Late Albian	*Neogastroplites* fauna
4.	Early Albian	*Gastroplites* fauna
	(Blairmore lower flora—pre-Albian)	
3.	Aptian	*Beaudanticeras affine—Deshayesites* fauna
2.	Late Barremian—early Aptian	"*Astarte*" *natosini* fauna
1.	Neocomian	*Aucella* fauna

Thompson and Axford (72, pp. 32, 42) list the following fossil succession in the Blairmore:

Haplophragmoides gigas
Gastroplites
Lemuroceras
Beaudanticeras

Inoceramus dowlingi
Ginko pleuripartita
Metacypris
Unio hamili

Bell (8, p. 517) describes the Blairmore as plant-bearing almost throughout, and as having two distinct floras: the lower, extending nearly throughout, begins 200 feet above the base and extends to about 300 feet from its top; the upper is confined to the top 300-foot interval. He considered the lower flora to be wholly Aptian and the upper flora to be early Albian. Bell (8, p. 519) shows that very few species of the early Blairmore flora are found in the Wealden of England (Wealden 4; see Fig. 2) and lists the following:

**Equisetum lyelli*	*Cladophlebis browniana*
**Onychiopsis psilotoides*	*Ptilophyllum dunkerianum*
Rufordia goepperti	*Elatides curvifolia*

**Ctenis borealis* may equal *Ctenis* sp.
Pseudocycas n.sp. close to *P. saportae* (Seward)

Except for those marked with an asterisk, this Wealden element is not found in the Kootenay. Bell (8, p. 519) concludes that if the Wealden of England is earlier than Aptian, and if the age of the lower flora of the Blairmore is judged solely on the basis of its Wealden element, it might be assigned a pre-Aptian age. The Wealden-Purbeck break in England has always been considered to be slight and, as shown in Fig. 2, the English Wealden is now known to be equivalent to the European Wealden 4 and is basal Cretaceous. A study of Fig. 2 indicates that the lower flora of the Blairmore is probably younger than the English Wealden. It is probably surprising that the lower flora of the Blairmore contains more of the characteristically Jurassic genera, e.g. *Sagenopteris*, *Zamites*, *Nilssonia* of the *orientalis* group, *Phoenicopsis*, and *Elatides*, than does the older flora of the Kootenay. However, it is the entrance of new species that has significance. An important newcomer in the lower flora of the Blairmore is *Elatocladus* (*Sequoia ?*) *smittiana* and also the first appearance of dicotyledon, *Sapindopsis variabilis* (8, p. 520). On the basis of flora, therefore, the base of the Blairmore would appear to mark an important time break and there appears to be no reason why the Blairmore does not range in age from basal Cretaceous to Aptian.

Important Considerations in Assessing the True Age of the Kootenay

European Chronology

Figure 2 shows the correlation of the Upper Jurassic and Lower Cretaceous of Europe with that of Alberta and British Columbia, and Montana and Wyoming. It will immediately be obvious that it is useless to speak of

the "Wealden" or the "Purbeckian" or "Valanginian," without being more specific. Speaking in general terms has resulted in considerable confusion in European chronology in the past, and is only now being clarified. Thus the Swiss Valanginian is only equal to the German and English Upper Valanginian; the English Middle Valanginian represents only the upper part of the German Middle Valanginian; the English Wealden represents a very small part of the German Wealden; and the Swiss Purbeckian is only equivalent to the lower part of the English Purbeck. Obviously, past comparisons with the "Wealden" flora and fauna will have to be re-examined.

Basal Blairmore Unconformity

It is probably easy now to look back and argue that the basal Blairmore unconformity is the logical Jurassic-Cretaceous boundary. However, there is not nearly enough importance placed on unconformities by most geologists. The unconformity at the base of the Blairmore was recognized by Rose (66, pp. 109–10) who first included the basal conglomerate in the Blairmore and described it as ten to thirty feet thick in valleys or channels eroded into the Kootenay. Stewart (71, p. 28) described the contact of the basal Blairmore conglomerate with the Kootenay as "conformable," showing that locally it may appear to be conformable and that the regional unconformable relationships can only be recognized by regional mapping.

Most geologists (12; 58; 20, p. 22; 23; 36, p. 17; 16, p. 66; 24, p. 144; 61; 62; 9, p. 2) are now agreed that the unconformity at the base of the Blairmore is regional in nature and truncates all pre-Blairmore rocks, ranging in age from Kootenay down through the marine Jurassic, Triassic, Permo-Pennsylvanian, Mississippian, Devonian, and older. In the Foothills and Canadian Rockies it is underlain by the Kootenay and is more difficult to recognize, particularly further west in the Fernie area, where the Kootenay becomes more conglomeratic at the top. Eastward, where it is underlain by the marine Jurassic, it is readily identified except locally where it is underlain by Jurassic sands. Beyond the eroded edge of the Jurassic, it is underlain by Mississippian and Devonian rocks and is quite apparent.

Basal Blairmore Conglomerate

The basal Blairmore conglomerate ranges from ten to sixty-five feet in thickness in outcrop sections. In the central Foothills it is named the Cadomin formation and in Montana and Wyoming the Kootenai or Cloverly conglomerate. The name Cadomin formation is here adopted for the mountain and Foothills area of Alberta and British Columbia. It is a massive resistant member composed of flattened and well-rounded pebbles of black, white, and green chert, white and grey quartzite, and quartz, varying from one-quarter inch to three inches in diameter. The pebbles are firmly cemented in a coarse sandy matrix.

Throughout the Plains area, the basal Blairmore is generally referred to as the basal quartz-sandstone, now named the Ellerslie member. As indicated

in Fig. 1, the Ellerslie extends far beyond the subcrop of known Jurassic sediments, overlying and filling depressions in an erosion surface composed of rocks of Palaeozoic age, and is overlain by strata of known Cretaceous age. Unquestionably, it belongs to the Cretaceous sequence and is post-Jurassic in age.

W. S. Yarwood (77) described the stratigraphy of the Spring Coulee well, forty miles south and west of Lethbridge, and clearly recognized the basal Cretaceous chert conglomerate and the characteristic *Unio hamili* zone of the Lower Blairmore. He correctly placed the immediately underlying Dalhousie sandstone in the Jurassic and assigned a Jurassic age to the coal at 5687 feet. His description is summarized as follows:

Lower Cretaceous	
4967′–5640′	Blairmore formation
at 5605′	Fish scales (*Cycloides*)
5625′	Shale, nonmarine, with *Unio hamili* and *Corbula onestae*?
5642′	Basal Cretaceous chert conglomerate
Upper Jurassic	
5640′–5667′	Brown ribbon sandstone (Dalhousie)
5667′–5911′	Dark grey shale
at 5687′	Coal and bentonite

Both the basal Blairmore unconformity and the overlying basal conglomerate, the Cadomin formation, and its equivalent in the Plains, the Ellerslie member, represent a regional hiatus separating the underlying rock sequences from the overlying Cretaceous sequence.

The Calcareous Member of the Lower Blairmore

The "Ostracod zone" has long been known to the petroleum geologist as a persistent and reliable marker in the Plains, occurring between the Glauconite sand member and the Ellerslie (Basal Quartz Sand) member and approximately a hundred feet above the unconformity (72). It is also recognized in the Foothills where it was referred to as the Calcareous member by MacKay (49, 50) and occurs 300 feet to 500 feet above the base of the Cadomin. In the Peace River region, the glauconite sand member is known as the Bluesky formation and is 35 feet to 150 feet thick—average: 50 feet. It lies immediately above the Ostracod zone.

The name Calcareous member was adopted by Glaister and is the same calcareous member of the lower Manville in Saskatchewan and the lower Kootenai of the Interior Plains of Montana. At Ryan Dam, Montana, it is 12.5 feet thick and is 152 feet above the unconformity at the base of the Kootenai (32, p. 620). It is thus an important and widespread marker.

The Calcareous member of the Lower Cretaceous was previously assigned

TABLE II

LOWER BLAIRMORE MICROFOSSIL ZONES*

Coaly sandstone member	230′	
Glauconite sand member (Home sand; Bluesky	255′	
Calcareous member (Ostracod zone) contains abundant *Metacypris persulcata* Peck *Metacypris angularis* *Metacypris pahasapensis* *Unio hamili*	0.5–75′	Wealden 4
Ellerslie (Basal Quartz sand) member	150′–350′	
REGIONAL UNCONFORMITY		
Kootenay, and older (Nikanassin)		

*Modified after Loranger, 45, 46, 47; Thompson and Axford, 72; Glaister, 32.

to the Aptian by Loranger (45) and to the Lower Albian by Badgley (3), but no supporting evidence is presented. Although the microfossils listed in Table II have been known to the petroleum geologist for many years, their significance has not been recognized for *Metacypris pahasapensis* is the index fossil of the European Wealden 4, and is basal Cretaceous. *M. pahasapensis* has been identified in several wells in Western Canada and this identification has been verified by J. B. Wolburg (Hubert Steghaus, personal communication). The unconformity at the base of the Blairmore is thus in reality the base of the Cretaceous. If the base of the Blairmore is Wealden 4 and the upper flora of the Blairmore is Albian, then the Blairmore represents the entire Lower Cretaceous, and is equivalent to all of the Neocomian (Berriasian, Valanginian, Hauterivian, Barremian), and the Aptian and Albian. The Kootenay (and Nikanassin) is now further limited by the basal Blairmore (basal Cretaceous) unconformity, which is now dated as basal Wealden 4. Thus the Kootenay is entirely late Upper Jurassic, and is equivalent to Portlandian 3 and all of the Purbeck beds of England, ranging from Portlandian 3 to Wealden 3. These relationships are illustrated in Figs. 1 and 2.

Fossil Vertebrate Evidence

It is interesting to note that fossil vertebrate evidence of the Jurassic in Canada is practically non-existent: the age of the greatest dinosaurs and earliest mammals is represented by fossil egg capsules, termed *Chimaerotheca*, of an early ratfish. These are from the Nikanassin formation exposed

on Sulphur River (42, p. 10) and were described by Warren (75) who considered them to be of Upper Jurassic age.

Re-evaluation of Kootenay Floral Evidence

The meritorious services of Bell, and the meticulous care with which he has prepared his evidence, are universally recognized. The evidence assembled has been interpreted to support a Cretaceous age for the Kootenay (8, 9), but can also be interpreted to support a Jurassic age. The Kootenay is distinguished from the Blairmore by the complete absence of dicotyledonous leaves in the Kootenay. Although the Kootenay flora is quite distinctive and demonstrates a late stage of evolution, there are definite affinities to Jurassic forms and many of the species do not range upward into the Blairmore.

"Most important of these are *Czekanowskia* cf. *rigida* and *Baiera* cf. *furcata*, both are macroscopically indistinguishable from the Jurassic species to which they are compared. So far as known, these two species do not occur in the Blairmore group of the type area of the Kootenay. A third Kootenay species worthy of note is *Nilssonia nigracollensis*, for *Nilssonia parvula* from the so-called Jurassic of Oregon is evidently conspicuous. Other species, except the long-ranging *Podozamites lanceolatus*, while allied to Jurassic species, are specifically distinct. A few aberrant specimens of *Cladophlebis virginiensis* are scarcely distinguishable from *Cladophlebis denticulata*, although the normal Kootenay forms have entire pinnules. *Ctenis borealis*, too, is seemingly very close to *Ctenis orovillensis* from the Oregon Jurassic, although considered to be specifically distinct.

"Those best defined Kootenay species that are not known as yet to range upward into the Blairmore group are *Sphenopteris cordai*, *Taeniopteris canmorensis*, *Baiera* cf. *furcata*, *Czekanowskia* cf. *rigida*, *Ptilophyllum robustum*, *P. hirtum*, *Nilssonia schaumburgensis*, *N. nigracollensis*, *Ctenis borealis*, *Pseudoctenis hazeltonensis*, *Pityospermum anthraciticum* and *Podozamites corbinensis*" (9, pp. 6–7).

"Although all but a few of the species occurring in the Kootenay flora range up into the lower flora of the Blairmore, the latter is marked by the arrival of forms not yet found in the Kootenay formation" (9, pp. 10–11). The most important of these are the Angiosperms, "which, although extremely rare, herald the flora that occurs in the upper part of the Blairmore at its type locality, in which dicotyledons predominate" (9, p. 11). The foregoing evidence suggests a considerable hiatus between the Blairmore and the Kootenay and is in agreement with the regional structural evidence.

Bell (9, p. 7) lists seven Canadian Kootenay species as occurring in the Kootenai of Montana. In view of the emended Kootenai of Brown (11), in which he recognizes the unconformity at the base of the Cretaceous which was formerly concealed within the old Kootenai, this list should be reviewed,

as some or all of these might be from the section below the basal Cretaceous unconformity, which is now included in the enlarged Morrison.

Age and Correlation of the Kootenai and Morrison of Montana

"In the Western Interior of the United States the establishment of the boundary between the Jurassic and the Cretaceous systems was long controversial . . . primarily because of a vigorous disagreement as to the age of the Morrison formation. There is now fairly general agreement that the Morrison is Jurassic and that the next overlying beds are Cretaceous" (15, p. 1013). In 1945, Cobban, and in 1946, Brown recognized an unconformity within the original Kootenai of Fisher (25, p. 50) and Calvert (13, p. 25) in Montana, and revised the stratigraphic section, placing the basal seventy-five feet of Kootenai in the Morrison. This was a step in the right direction and the emended Kootenai is now entirely Cretaceous and essentially the equivalent of the Blairmore to the north and the Cloverly to the south.

No assured sedimentary record has yet been recognized for that part of Early Cretaceous time that corresponds to the Neocomian stage of the European sequence (15, p. 1022). However, the Kootenai formation of Montana contains a non-marine fauna and flora like that of the lower part of the Blairmore formation in southern Alberta and British Columbia, and the Gething and McMurray formations in northern British Columbia and Alberta (58). From experience in Canada, it can now be assumed that the Neocomian is represented by the lower Kootenai (and Cloverly), and this will probably be demonstrated before long by the use of microflora.

The succeeding interval in the Early Cretaceous, corresponding to the Aptian stage of the European sequence, is represented by the (upper) Kootenai of Montana, based entirely on the relationship of the flora to those of other regions (15, p. 1022).

Cobban and Brown's enlarged Morrison is now equivalent, at least in part, to the Canadian Kootenay and the underlying Passage beds, and is entirely Upper Jurassic. Based on *Chara* and Ostracods, Cobban (14, p. 1269) placed the lower part in the "Kimmeridgian and possibly younger," which is in agreement with earlier tentative dating by Fisher (25, p. 30) and Calvert (13, p. 24). In 1957, Francis (26, fig. 13, p. 394) showed the northern truncated edge of the Morrison formation extending along the Forty-ninth Parallel and lapping up into southern Alberta, southwestern Saskatchewan, and southwestern Manitoba. He placed the Morrison in the Jurassic and as equivalent to all of the Kimmeridgian and part of the Portlandian, the Purbeckian and some of the Portlandian having been removed by erosion.

Imlay (35, Chart 8C, p. 953) shows the Morrison of Montana as equivalent to the Kimmeridgian and the lower part of the Portlandian, and states (p. 953) that the exact upper limit is not known. The underlying marine

beds carry Oxfordian ammonites, and as per Imlay, Simpson (68) correlated the reptiles with the Portlandian of East Africa, and Jones (37) correlated Ostracods with the Purbeck beds of England.

The age of the Morrison and the Kootenai of Montana and their correlation with the Canadian Kootenay and the Blairmore is shown in Fig. 2. The top of the Morrison is an erosion surface, controlled by pre-Kootenai erosion which has removed all the Morrison and has cut down into the underlying Ellis group over the Sweetgrass Arch (14). As the basal Kootenai is Wealden 4, the Morrison is entirely Jurassic and may range up to Wealden 3.

Age and Correlation of the Mattagami Formation, James Bay Lowland, Ontario

Two areas of Mesozoic strata in the James Bay lowland were named the Mattagami formation by Keele (38, p. 47), who with M. Y. Williams made the first collection of fossil plants from these beds. The Mattagami formation has received further study by McLearn (55), Bell (7), Dyer (22), and Martison (53). The early collections of fossil plants were identified as not younger than Kootenay and possibly older. Later, the collections were restudied by Bell (22) who considered the assemblage to be either Upper Jurassic or Lower Cretaceous, giving preference to the latter because of the presence of *Pityophyllum graminaefolium* which is widespread in the Kootenay.

Now that the Kootenay has been shown to be Upper Jurassic in age, the Mattagami formation can be placed in the Upper Jurassic with confidence.

Hazelton Group of British Columbia, and Tantalus Formation of the Southern Yukon Territory

Fossil plants from the upper part of the Hazelton group have been assigned a Lower Cretaceous age based on correlation with the Kootenay flora (9, p. 23). It was also pointed out that some species resemble common Jurassic forms. Plant fossils collected from the Tantalus formation were tentatively considered to be Neocomian in age, with the possibility of a late Jurassic age (9, p. 27).

Radforth and Rouse (65, p. 21) state: "The number of microfossils common to the Kootenay and Tantalus indicates that these two formations contain similar floras, supports the contention that the two horizons are correlative, and suggests a uniformity of the flora over a wide geographic region."

As the Kootenay can now be assigned an Upper Jurassic age, both the Hazelton group and the Tantalus formation (as well as a portion of the Jackass Mountain group; (9, p. 30) can now be placed in the Upper Jurassic, thus removing the supposed conflict between fossil plant and invertebrate fossil evidence (2; 9, p. 24).

Conclusions

The age of the Canadian Kootenay is sharply restricted, ranging from Portlandian 3 for the Moose Mountain member, or basal Kootenay sandstone, to not younger than Wealden 3. It is thus mainly equivalent in age to the Purbeck beds of England and is uppermost Upper Jurassic.

As a result of placing the Kootenay in the Cretaceous and correlating it with the Barremian, many false conclusions and assumptions have been made and should be revised by those competent to do so. This will probably also require a revision of our present ideas of evolution of fossil plants.

Probably of equal importance is the fact that the new Canadian evidence supports the European correlations proposed by Bartenstein and Burri (4). The regional unconformity at the base of the Cretaceous, which is a continental regional unconformity in North America, emphasizes the importance of the pre-Wealden 4 unconformity in Europe.

The new age designation of the Kootenay will require many revisions in published maps and reports, and in our understanding of the fossil record of the Upper Jurassic and Lower Cretaceous. It demonstrates that the only real time-break is the obvious well-known unconformity at the base of the Blairmore.

The importance of unconformities in working out the geological record cannot be overemphasized. As pointed out by Bell (8, p. 514), "once detected it destroys the unity of the formation in which it was included, and it must become the boundary between separate formations." Of equal importance are the unconformities between rock sequences or systems. As these are regional in nature, and as the limiting beds will vary in age from close to far apart, they cannot always be recognized in the beginning, and frequently must await regional mapping. Once they become apparent, their importance must be acknowledged and the overlying and underlying sequences must be sharply separated. The "Dunlevy formation" (6) is an excellent example of this. It is now obsolete and must be broken down into the Cadomin formation (Lower Cretaceous, early Neocomian) and the Nikanassin (Uppermost Jurassic).

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TRANSACTIONS OF THE ROYAL SOCIETY OF CANADA
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SECTION FOUR

On the Origin of Algal-like Forms and Carbon in the Sudbury Basin, Ontario

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ABSTRACT

The sulphide bodies at Consolidated Sudbury Basin Mines occur in a thinly laminated limestone that contains algal-like forms. This limestone is a member of the Vermilion formation which overlies the glowing avalanche deposits of the Onaping formation and underlies the Onwatin-Chelmsford sedimentary formation. In addition to algal-like forms, oolites, pisolites, and evaporite breccias are present. The similarity of lithology, structure and geological environment of the thinly laminated limestone and associated chert-breccia to travertine that is forming today in the hot springs of Yellowstone National Park, Wyoming, is pointed out.

Carbon, in the form of graphite, is widely distributed throughout the sedimentary rocks of the Sudbury basin and is locally concentrated in veins of anthraxolite. These carbonaceous sediments are about the same age (1700–1800 m.y.) as the Gunflint iron formation near Schreiber, Ontario, in which structurally preserved plant forms have been described. It is concluded that the evidence favours a hot-springs origin for the limestone containing the algal-like forms and that the carbon is of organic origin.

THE origin of the carbon in the rocks of the Whitewater group in the Sudbury basin has long been the subject of speculation by geologists. Coleman (3, p. 100) thought that the carbon content of the Onwatin slate and associated anthraxolite was due to the presence of organic material in the sediments. Burrows and Rickaby (2) agreed, but suggested that the carbonaceous material in the Onaping tuff might be formed by reduction of CO_2 and CO in gaseous emanations attending volcanism. Williams (15, p. 82) doubted if this could happen and thought it best to consider the carbon of both the Onaping and the Onwatin formations as organic in origin. He suggested the possibility of glowing avalanches being discharged into a water-filled basin containing organic matter. Subsequent circulation of ground waters could have carried carbonaceous materials downward into the tuff.

Martin (7) first recognized that the sulphide mineralization at Consolidated Sudbury Basin Mines is confined to a stratigraphic unit which he called the Vermilion formation. In discussing drilling results he stated (7, p. 36): "Using the chert as a guide, the strictly stratigraphic relationship of the ore horizon to other formations became apparent and was soon confirmed by underground study." He described in the Vermilion formation a carbonate member which forms the host rock of the ore. He also noted the

sedimentary character of this member and associated chert and chert-breccia, the presence of oolites and pisolites, the fine wavy banding such as might result from colloidal deposition and concluded that ". . . some, or most, of the chert, carbonate rock, and pyrite may be of hot springs origin."

Williams (15) and the author (10) interpreted the Sudbury basin as a subsidence structure associated with Pelean-type volcanism. It was envisioned that, within the resulting closed stagnant basin, hot springs emanating from the glowing avalanche deposits could produce an environment in which organisms such as algae and bacteria could exist. Thus an accumulation of carbonaceous mud intermingled with limestone and chert of hot-springs origin could be laid down. The similarity of the geological environment within the Sudbury basin to that of areas of recent volcanic activity with associated hot-springs deposits was noted. Later, the author had the opportunity to visit Yellowstone Park where travertines of hot-springs origin are now being deposited upon Pelean-type volcanic rocks. Analogous features of recent hot-springs deposits to the chert and algal-like carbonates of Consolidated Sudbury Basin Mines were noted and are reported in this paper.

Geology of the Area

The geology of the Sudbury basin has been described recently by Thomson (10) and Williams (15) and the property of Consolidated Sudbury Basin Mines by Martin (7). The company owns two main mines, the Errington and the Vermilion. Although they are about four miles apart they are located at the same stratographic horizon and their geology is quite similar. It should be pointed out that the Vermilion formation, which contains the algal-like forms, does not outcrop anywhere in the Sudbury basin, except for one small isolated exposure near the No. 1 shaft of the Errington mine. Another noteworthy feature is that the mines lie within a zone of great structural disturbance. The rock formations within the mine workings are intricately sheared, faulted, altered, and mineralized. This naturally makes it more difficult to recognize and interpret primary structures in the rocks.

The stratigraphic succession in the vicinity of the Errington and Vermilion mines is given in Table I. The areal extent of the Vermilion formation beyond the mine workings is unknown. Martin (7, p. 367) reports that "a fair amount of exploration drilling was done outside the mine areas and a considerable amount of barren chert and carbonate [was] found." The intervening ground between the Errington and Vermilion mines was drilled and numerous intersections of the argillitic member of the Vermilion formation were reported (10, p. 55), but only occasional limy bands were found. There is no record that drilling has located the Vermilion formation elsewhere in the basin. Thus, to date, there is nothing to indicate that the limestone and chert of the Vermilion formation is a very extensive unit, and it could consist of a number of local, disconnected lenses.

TABLE I

THE STRATIGRAPHIC SUCCESSION AT THE ERRINGTON AND VERMILION MINES

Formation	Description	Thickness
Chelmsford-Onwatin formation	Thinly bedded carbonaceous slate derived from mudstone and siltstone. It contains seams of nodular pyrite along bedding planes (possibly syngenetic). There is a transition upwards into carbon-bearing greywacke, argillite and arkose containing numerous concretions.	Not measurable but of considerable thickness
Vermilion formation	*Argillite and limestone*: finely bedded argillite with interbeds of limestone and dolomite. Regular beds up to 6 inches in thickness.	20 feet
	Chert-breccia: black chert fragments in a white chert matrix. Some algal-like banding and oolites.	Average 10 feet up to 20 feet
	Cherty carbonate: transition zone between thinly laminated limestone and chert breccia	Averages 10 feet Up to 50 feet
	Thinly laminated limestone: the sulphide-bearing horizon. It contains algal-like forms, oolites, pisolites, mamillary and crustiform banding.	Up to 100 feet
	Basal argillite: black, dense, siliceous mudstone, grading into tuff. Possibly a leached volcanic mud.	0–100 feet
Onaping formation	Fine-grained carbonaceous tuff at the top, grading downward into andesitic tuff, coarse tuff-breccia, rhyolite and rhyolite breccia. These are glowing avalanche deposits and dyke-feeders of Pelean-type volcanism.	Not measurable but in the order of thousands of feet

ALGAL-LIKE FORMS AND PISOLITES

The author did not recognize the algal-like forms at Consolidated Sudbury Basin Mines until after a visit to Yellowstone Park in 1957. Underground operations at the mine had then ceased and it was only possible to get specimens of these structures by examining the ore dump at both the Errington and the Vermilion mines. Naturally most of the rock on these dumps consists of sulphides, but occasional large specimens of sulphide-free bedded carbonate rock may be found. Several of these were collected from each of the mines and examined.

Typical algal-like forms are thinly laminated, the beds varying from paper-thin to about a tenth of an inch in thickness. Delicate wavy banding is characteristic of the rock. These tiny crenulations are subsidiary to crinkling on a larger scale that produces mamillary shaped surfaces with "domes and basins." Convex mamillary forms up to several inches in diameter were found. Generally the rock is so deformed that any larger structures would have been destroyed. The wavy lamination is accentuated by an alternation of dark- and light-coloured beds. The dark colour is due to the presence of innumerable tiny flakes of graphite. The light-coloured beds are crystalline carbonate, mostly calcite. A small amount of primary

silica was seen in the thin sections examined and a few grains of sulphide are generally present.

Pisolites and oolites are distributed throughout the algal-like forms. Some of these show very fine concentric banding with alternation of light- and dark-coloured rings. In some cases the larger pisolites have a thick and irregular outer coating. Some of the granules are composite or are clustered in small groups. Graphite occurs as extremely thin concentric rings in the oolites and pisolites and as interstitial filling between granules. Rows of partially developed pisolites have produced the wavy banding in places and may be considered as pseudo-pisolitic forms and crusts. Generally the crusts are nearly flat and essentially parallel to the bedding, but occasionally they become well arched to the hemispherical or cabbage-like forms so typical of good algal structures. The type of algal forms resembling thimbles stacked in vertical columns was not observed. Under the microscope the only structures visible are a lamination parallel to the surface of the algal-like form. This is recognized by the extremely thin interbeds of graphite flakes. The carbonate material is completely recrystallized so that its original texture has disappeared.

Several specimens of algal-like limestone were found that show cross-

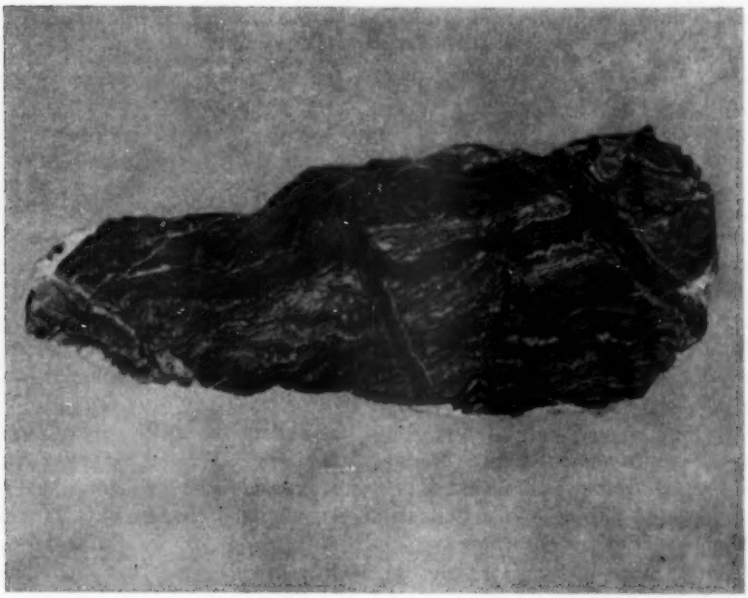

FIGURE 1. Polished surface of thinly-laminated limestone with algal-like forms. This was obtained from the ore dump at the Errington mine. The specimen shows oolites, pisolites, and a set of beds that crosscut the wavy crenulation of the algal-like beds. Tiny veinlets of later quartz and calcite may be seen.

cutting relationships of one kind of bedding to another. This structure is illustrated in the polished slab of limestone shown in Figure 1. Narrow beds of thinly bedded limestone cut directly across the normal wavy bedded variety. Sometimes these crosscutting beds will pinch out, but the line or zone of shattering will continue for some distance through the wavy beds. In places the crosscutting beds will terminate abruptly and merge with similar beds in the wavy sequence. In such cases the two directions of bedding may form almost a right angle.

Most of the structures described above may be seen today in the travertine deposits at Mammoth Hot Springs in Yellowstone Park. Here the hot waters have built up the well-known travertine terraces. The deposits when freshly formed are often brilliantly coloured by algae that live in the hot waters. This plant life is the principal agent in the formation of travertine. In the Mammoth travertine deposits thinly laminated mamillary forms produced by algae are well developed, and oolites and pisolites are associated with these structures. The rapid change in the attitude of beds is a characteristic feature of hot-spring deposits. Thinly laminated flat beds are laid down in the terrace pools and at the same time near-vertical beds are formed by precipitation of carbonates from hot waters flowing over the edge of the terraces. These steep beds will then be enclosed by

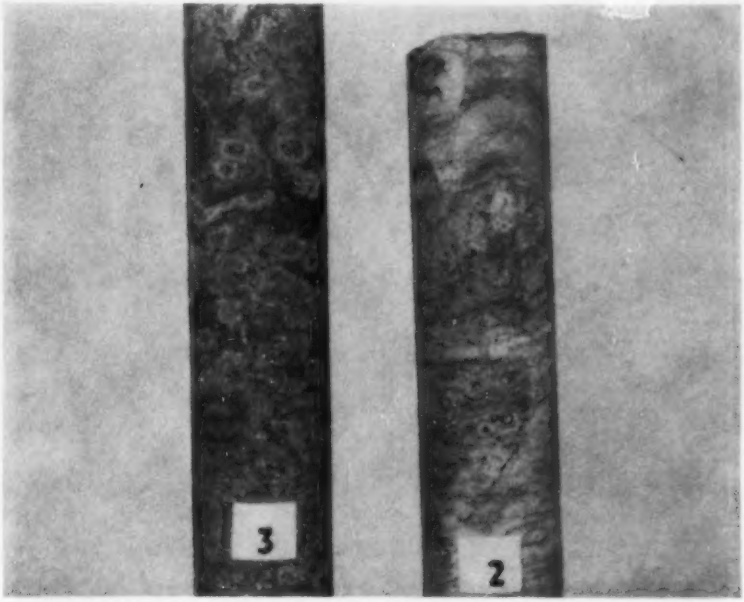

FIGURE 2. Drill core specimens of the host rock of the sulphides in the Errington mine. No. 2 shows the thinly laminated limestone with wavy bedding and oolites. No. 3 shows the development of pisolites in limestone.

flat beds built up by waters escaping by another route. In addition, drying out of the travertine terraces produces fissures along which hot waters circulate. Thin layers of carbonate are precipitated along the walls of the openings. The resulting rock may be classified as an evaporite breccia. Thus, the mechanism of travertine deposition at Yellowstone Park could readily explain the bedding relationships that are found in the thinly laminated limestone of the Sudbury basin.

Cherty carbonate and chert-breccia lie immediately above the algal-like limestone in the Errington and Vermilion mines. The chert-breccia consists of dark-coloured chert in a matrix of lighter-coloured chert and could result from shrinkage of chert beds due to evaporation and subsequent recementation. In other words, this, too, could be an evaporite breccia. In 1957, the writer was shown identical chert-breccia interbedded with travertine on the property of Montana Phosphate Products near Garrison, Montana.

It is important to note that the structures in limestone and chert-breccia of the Sudbury basin are primary depositional features and not the result of secondary tectonic deformation. Brecciation, fracturing, vein-filling, and mineralization, all associated with later deformation, have been superimposed on the primary structures described above, but it is easy to distin-

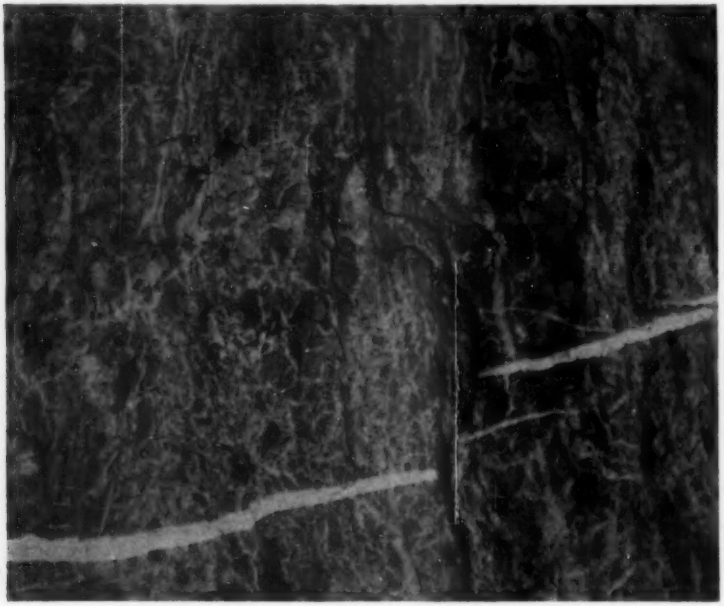

FIGURE 3. Chert-breccia on the drift wall, 300-foot level, Errington mine. The sulphide mineralization occurs in thinly laminated limestone that lies stratigraphically a few feet below this breccia.

guish betwen depositional and tectonic structures. No doubt much of the later quartz and carbonate veining is the result of regeneration of the original limestone and chert beds during deformation and metamorphism. Also, veinlets of quartz and anthraxolite are found in the algal-like limestone in the Vermilion mine. The anthraxolite is identical to that in the well-known occurrences located a short distance north of the mines.

It should be pointed out that there are limestones in the mines of a different type from those described above. These have regular beds from one to six inches thick. They show no evidence of algal-like structures or pisolites. This type of limestone lies stratigraphically above the thinly laminated limestone and chert-breccia. It is interbedded with buff-coloured argillite which is finely laminated in places. This could be a lacustrine detrital sediment, somewhat akin to present-day marl, and derived from the disintegration and reworking of the underlying algal-like limestone, chert, and volcanic muds.

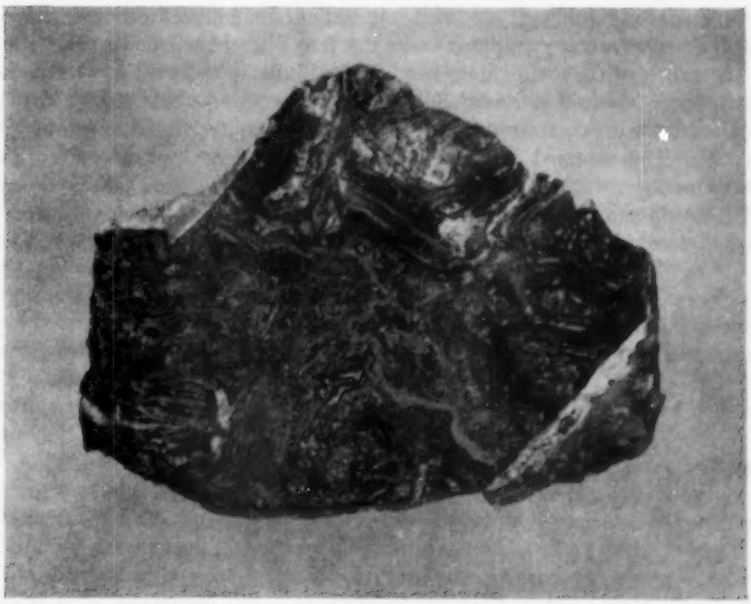

FIGURE 4. Polished surface of thinly laminated limestone that has been brecciated and mineralized with veinlets of sulphides. Later quartz-carbonate veinlets are white in colour. Specimen from ore dump, Errington mine.

CARBONACEOUS SEDIMENTS, ANTHRAXOLITE, AND CONCRETIONS

Coleman (3, p. 102) reported analyses of 6.8 to 10 per cent carbon in the Onwatin slate. Burrows and Rickaby (2, p. 28) estimated that the average

carbon content of the slate would be less than 1 per cent; analyses obtained by them ranged from 0.30 to 3.35 per cent carbon. Anthraxolite occurs with quartz and pyrite in fissure veins in the slate and small amounts are found within the ore zone in the Vermilion mine. The average composition of anthraxolite at the main occurrence (lot 10, concession I, Balfour township) is given at about 74 per cent carbon, 20.5 per cent ash, 4 per cent moisture, and 1.30 per cent volatile matter. Coleman believed (3, p. 103) that this anthraxolite was derived from the adjacent carbonaceous slate. The Chelmsford formation, which lies above the slate, also contains a small amount of carbon which accounts for its dark colour. The upper part of the Onaping formation contains considerable carbonaceous material. All the members of the Vermilion formation contain carbon in the form of graphite and two analyses given in Table II indicate 0.47 and 3.15 per cent carbon. In thin section this graphite is seen to occur largely as tiny flakes along bedding planes although an occasional graphite veinlet is also found. The field and microscopic evidence strongly favour a syngenetic origin for the carbon although much of it was later remobilized along slips and faults. Several strong graphite faults are found in the mine workings.

A feature of the sedimentary rocks in the Sudbury basin is the widespread occurrence of ovoid to rounded concretions averaging about a foot in diameter. Cylindrical concretions have been found in the Errington mine (2, p. 36). The rounded concretions are richer in lime and iron than the surrounding rocks. Concretions and carbon-bearing strata do no occur in the sedimentary formations outside the Sudbury basin. Their occurrence is therefore related to local environmental conditions within the basin.

TABLE II

ANALYSES OF ALGAL-LIKE LIMESTONE

Locations*	No. 1	No. 2	No. 3	No. 4	No. 5
Insoluble (in HCl)		7.56			
SiO_2	0.25	1.84		16.46	25.07
Al_2O_3	0.24				
Fe_2O_3	2.54				
FeO	4.56				
S	2.20				
MgO	15.27	4.14	11.10	0.88	1.20
CaO	27.80	40.92	27.42	41.05	30.29
H_2O+	0.34				
H_2O-	0.02				
CO_2	41.67	36.34	37.80	29.78	27.08
C	0.47	3.15			
MnO	1.21				
Zn	3.83				
TiO_2	Trace				
P_2O_5	Trace				
Total	100.40	93.95	76.32	88.17	83.64

*No. 1. Other half of polished slab shown in Plate I. Ore dump, Errington mine. No. 2. Thinly laminated limestone. Ore dump, Errington mine. No. 3. Bedded carbonate rock. Errington mine. No. 4. Limestone, 750-foot level, Vermilion mine. No. 5. Limestone, drill hole No. 197, Errington mine.

Other Research Methods

In recent years laboratory methods of attempting to establish the organic or inorganic origin of carbon in rocks have been developed. These are (*a*) the ratio of C^{12}/C^{13} isotopes in carbon-bearing material, and (*b*) biochemical analyses which detect remnants of amino-acids in fossils and carbonaceous sediments (1). Abelson (personal communication) has examined anthraxolite from the type locality in the Sudbury basin but has been unable to find any extractable compounds for amino-acid tests. The author has not heard of any carbon isotope ratio determinations on the rocks of the Sudbury basin. However, Craig (4) has questioned the reliability of the method and it is doubtful if such research would give a definite answer to the origin of the carbon.

Geological Environment of the Sudbury Basin

Volcanism and sedimentation in the Sudbury basin have been interpreted by Thomson and Williams (11). Their conclusions are slightly re-interpreted and summarized here as follows:

1. Rapid eruption of glowing avalanches in great volume (Onaping formation) with accompanying central subsidence formed a volcanic-tectonic sink or depression.

2. With the cessation of volcanism hot-spring deposits emanating from the Onaping pyroclastics produced algal-like limestone and chert locally within the depression. Simultaneously black carbonaceous muds were accumulated in stagnant shallow water in the lagoon-like tectonic sink by erosion of the adjacent tuffs and volcanic mud flows. Lowly organisms, such as algae and bacteria, could thrive in such an environment and account for the high carbon content of the sedimentary rocks. According to Stanton (9, p. 26) the conditions required for heavy accumulation of organic matter in this environment are high concentration of organisms, shallow water, medium rate of detrital sedimentation, and fine particle size.

3. The sulphide deposits of the basin could have been formed by fumarolic and hydrothermal leaching of the Onaping pyroclastics and precipitated by anaerobic bacterial activity in the reducing environment of the black carbonaceous muds in the manner postulated by Stanton (9).

4. The carbonaceous muds were gradually covered by coarse detrital material (Chelmsford arkose). This was derived only in part from the Onaping pyroclastics and mainly from a plutonic and metamorphic terrain. The basin was still supplied with organic material, and carbonate concretions were formed in abundance. Weeks (14, p. 100) has shown that carbonate concretions are characteristic of a stagnant-water environment in which oxygen is quickly used up, carbon dioxide accumulates from organic sources and lime is retained in solution as bicarbonate. Such concretions are of syngenetic or diagenetic origin.

Age of the Sedimentary Formations in the Sudbury Basin

Various interpretations of the age and correlation of the formations in the Sudbury basin (Whitewater group) have been made. It is not certain whether they should be classified as early or late Precambrian in the geological time scale, although the author believes that the evidence favours an early Precambrian age (10, pp. 25–9). Isotopic dating of the Sudbury irruptive which intrudes the Whitewater group (8, p. 55) gives ages from 1600 to 1800 million years, and the best age determination of rhyolite in the Onaping formation is 1700 million years (H. W. Fairbairn, personal communication). The age of the algal-like forms and carbonaceous sediments might be tentatively placed at about 1700 million years, although it would be in excess of 1800 million years if the maximum age determination is used.

Summary and Conclusions

The writer is aware of the caution that is necessary in assessing algal-like forms and the difficulty of establishing the organic origin of carbon in ancient sedimentary rocks (5; 6; 16). No claims for the discovery of positive evidence of organisms in the Sudbury basin are made here and only certain analogies are pointed out. In the writer's opinion the similarity of the thinly laminated limestone and chert-breccia to travertine and chert deposits of Yellowstone Park and vicinity would strongly suggest a hot-springs origin of the Sudbury rocks and that algae aided in their formation. The delicate laminate of carbon flakes along bedding planes cannot logically be explained in any other way than by the entombment of organic material in the process of sedimentation. In fact, the wide and uniform distribution of carbon throughout all sedimentary formations inside the Sudbury basin, in contrast to the lack of carbon in sedimentary rocks outside the basin, can only be explained by an abundance of organisms within a coastal lagoon-like body of stagnant water. The introduction of carbon from igneous sources seems to be out of the question, especially in view of the fact that it is very doubtful if volcanism can produce carbon in any quantity (15, p. 82) and the basin is generally devoid of intrusives. Therefore, the field relations, even when unsupported by laboratory tests or fossil evidence, would almost certainly indicate an organic origin for the carbon.

It should be remembered that structurally preserved plant forms have been found in the black chert of the Gunflint iron formation north of Lake Superior near Schreiber (12). This chert shows a consistent age of about 1700 million years by the potassium-argon method of dating (13, p. 1304). The age of the black carbonaceous sediments in the Sudbury basin is in the range of 1700 to 1800 million years. It is, therefore, not at all unreasonable to believe that organisms or plants could have existed there at the time of sedimentation, especially in view of the favourable geological environment. S. A. Tyler has informed the author that the plant forms in the Gunflint

formation have been found at only one horizon, two inches to three feet in thickness. There is a possibility that fossil remnants of microscopic dimension exist in the Sudbury basin, but the chances of discovery are very small if their distribution is on the same scale as in the Gunflint formation.

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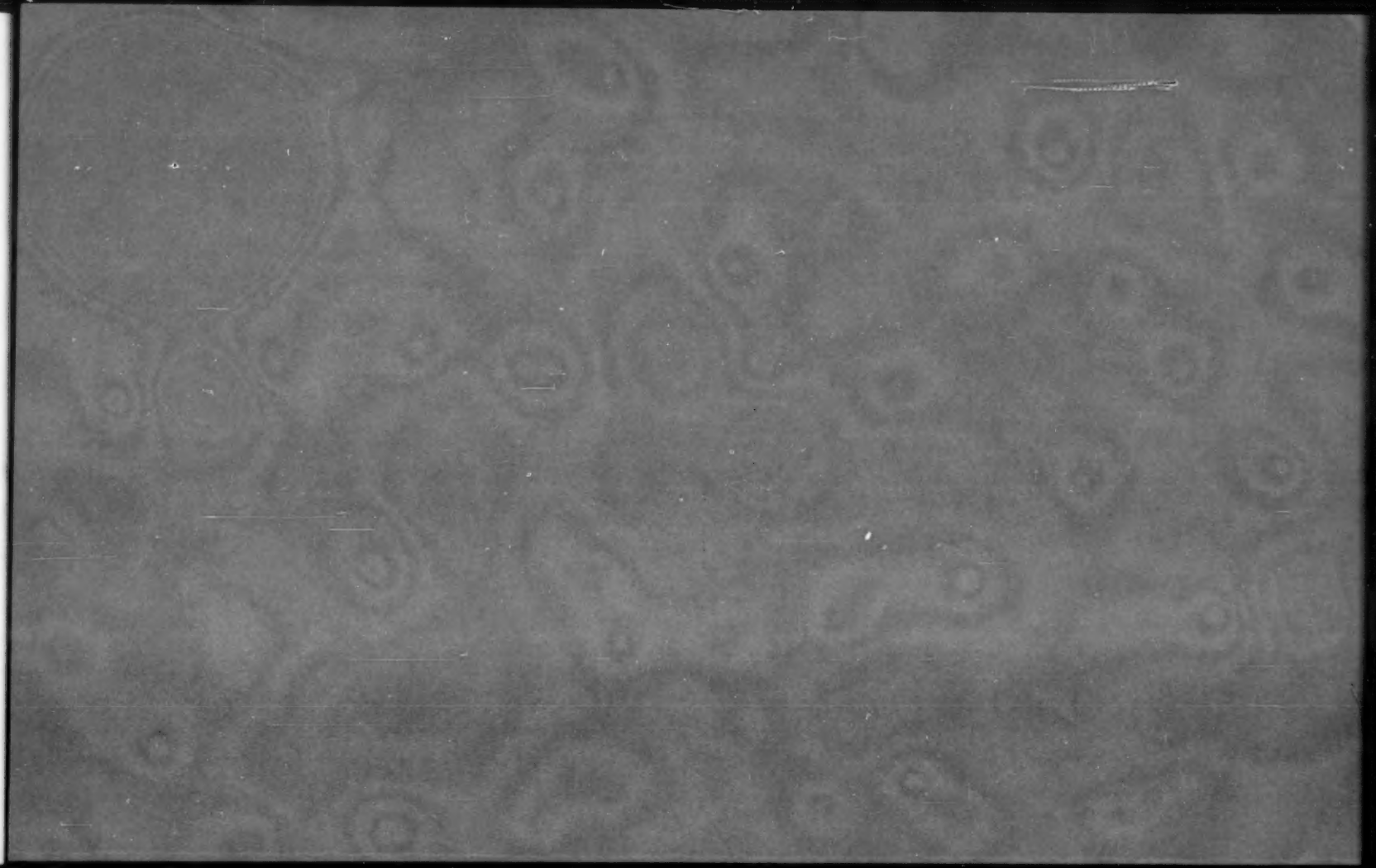

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